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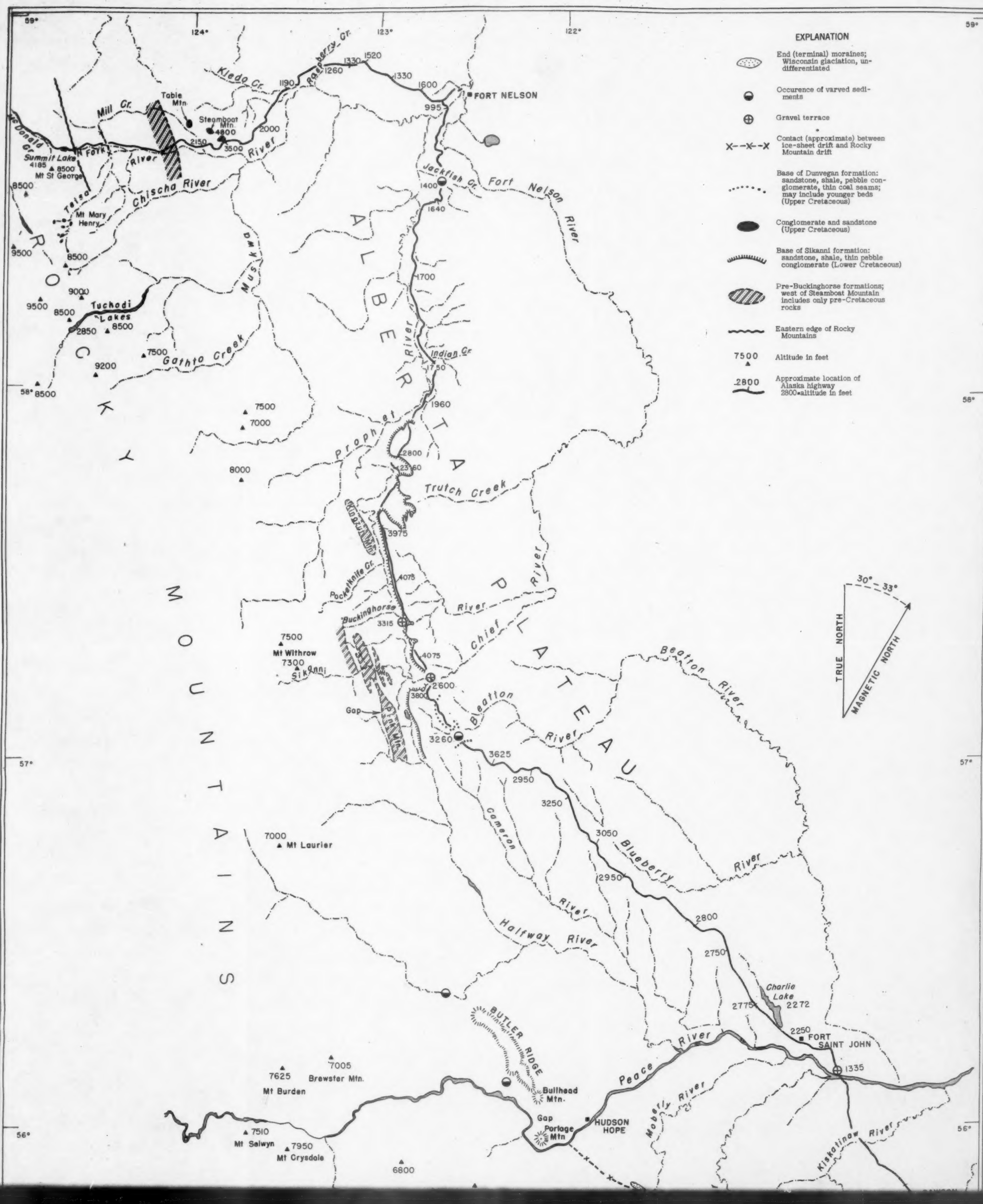
59°
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56°
EXPLANATION
End (terminal) moraines; Wisconsin glaciation, undifferentiated
Occurence of varved sediments
Gravel terrace
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Kledo Cr.
Mill Cr.
Table Mtn.
Steamboat Mtn. 4800
McDonald Cr.
Summit Lake 4185
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N Fork
Fork River
Chischa River
Talsa
Mt Mary Henry
Muskwa River
Tuchodi Lakes
Tuchodi
Gathto Creek
Jackfish Cr.
Fort Nelson River
Indian Cr.
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Pocketknife Cr.
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30° - 33°
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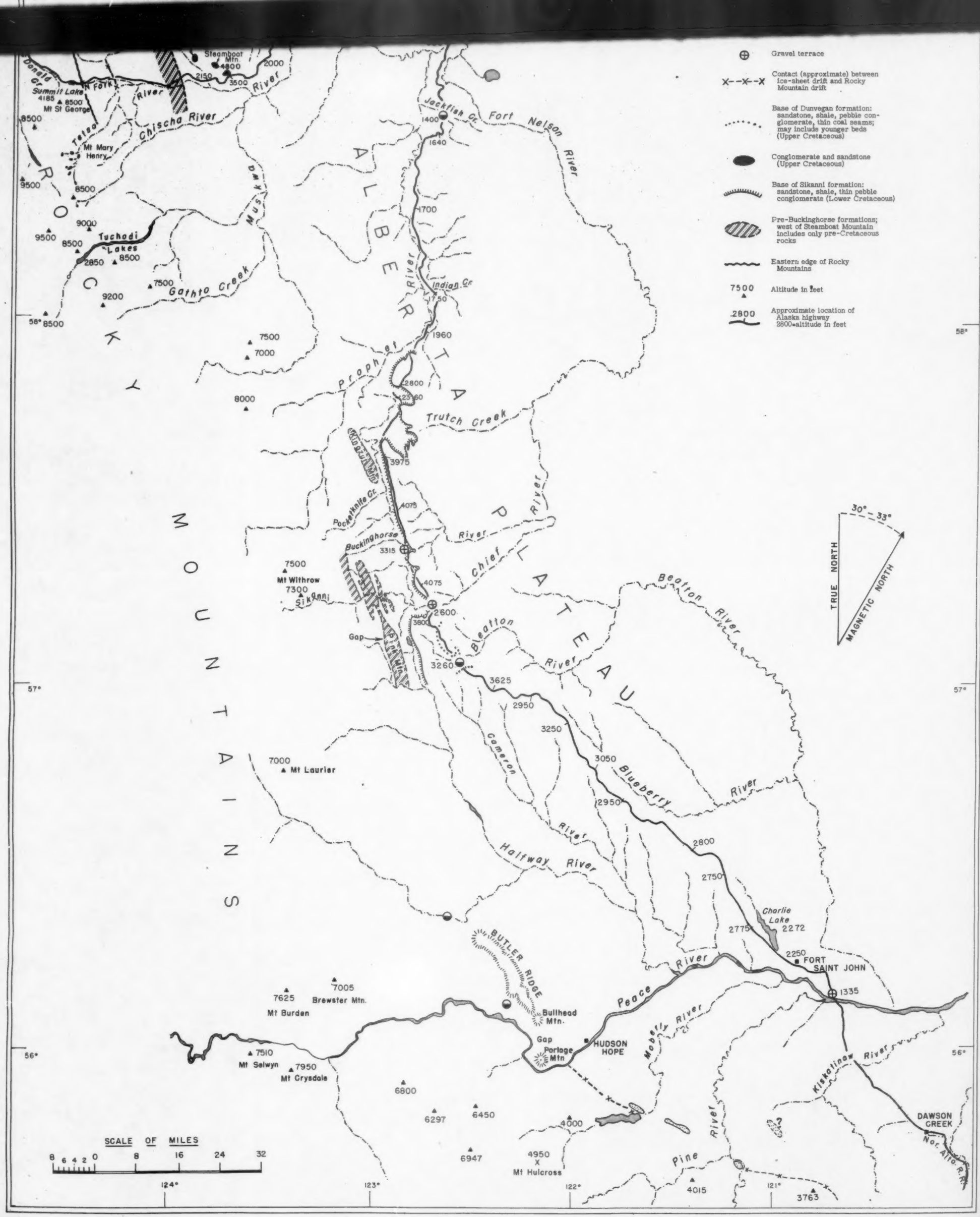
MAGNETIC NORTH



MAP OF AREA FROM DAWSON CREEK TO FORT NELSON, BRITISH COLUMBIA

Base map and formation boundaries from preliminary map 44-30A and 44-28A, Geol. Survey Canada; and sheets 93 N.E., 94 N.W., 94 S.E., Hydrographic and Map Service, Surveys and Engineering Branch, Canada Dept. Mines and Resources. Data on glacial geology from Hage (1944), Williams (1944), Flint and others (1945), and field work along highway by Denny.

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LATE QUATERNARY GEOLOGY AND FROST PHENOMENA ALONG ALASKA HIGHWAY, NORTHERN BRITISH COLUMBIA AND SOUTHEASTERN YUKON

By Charles S. Denny

Abstract

Reconnaissance field work along the Alaska Highway in northern British Columbia and southeastern Yukon furnishes preliminary data on the later Quaternary history of the region, and on the processes and results of intensive frost action. Extensive erosion surfaces were developed prior to glaciation, such as the Alberta Plateau of northeastern British Columbia and the Yukon Plateau in southern Yukon.

In the region from Dawson Creek to Fort Nelson, British Columbia, the drift is dominantly a slightly weathered bouldery clay till of Wisconsin age, deposited by ice that came from the east. In the northern Rocky Mountains, the last eastward advance of the Wisconsin glaciers to the mountain front preceded the maximum westward advance of the ice sheets. In the foothills belt west of the Highway, the tills of these two advances are separated by lake deposits. Much of the drift is probably pre-Altamont in age and was subjected to vigorous frost action during the Altamont substage. The climate ameliorated perhaps with the advent of the post-glacial optimum and possibly coincident with the development of prairies. A recent change from prairie to forest perhaps indicates a slight cooling in recent time. In the northern Rocky Mountains, four substages of Wisconsin glaciation are recognized tentatively on the basis of morainal deposits and outwash terraces. The upper Liard basin is characterized by extensive pitted outwash plains, probably of late Wisconsin age. From Teslin Lake to Whitehorse, the Highway is bordered by extensive terraces of gravel and sand deposited in ice-marginal streams and lakes.

Intensive frost action has modified pre-existing land forms and has produced a wide variety of features, such as talus, blockfields, stone rings, stone stripes, and terraces. Long smooth slopes, chiefly due to mass movements such as solifluction, are characteristic of the landscapes. These slopes and ancient soil structures extend down into forested areas and probably developed prior to the advent of forests, doubtless prior to the post-glacial optimum. Phenomena resulting from intensive frost action at the present time are restricted largely to areas above timber line.

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INTRODUCTION

The Alaska Highway

The Alaska Highway extends for about 1422 miles from Dawson Creek, British Columbia, to Big Delta, Alaska (Fig. 1), where it meets the Richardson Highway that connects Valdez and Fairbanks, Alaska. Its construction, authorized in the winter of 1942, was to be "along the line of the then existing chain of airports from Edmonton, Alberta, to Fairbanks, Alaska, fur-

nishing a supply route to Alaska and, at the same time, servicing and supplying the airfields along the route and providing a means of safety for personnel engaged in ferrying aircraft from the United States to Alaska" (Scrugham, 1943). The Army "pioneer road" was bulldozed by U. S. Army Engineers between March and November 1942; the final highway with gravelled surface, which closely follows the pioneer trail, was built by Canadian and American contractors under the supervision of the Corps of Engineers and the U. S. Public Roads Administration (Robinson, 1946). The Canadian portion of the Highway was turned over to the Canadian government in 1946.

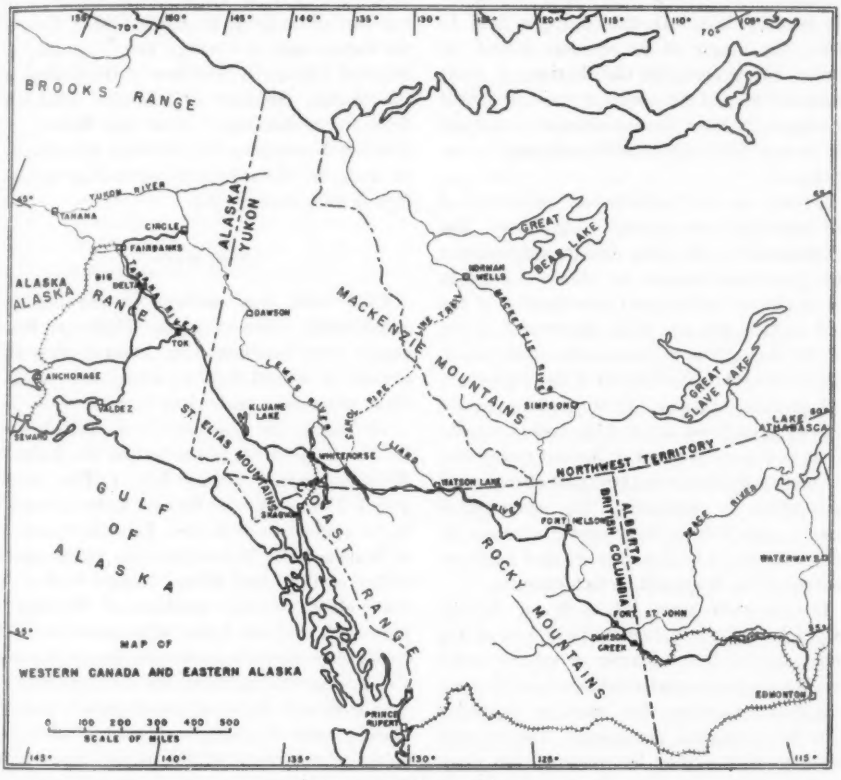


FIGURE 1. INDEX MAP

Purpose and Scope of Study

In 1943 the author made a reconnaissance trip over the southern part of the Alaska Highway from Dawson Creek, British Columbia, to Whitehorse, Yukon. The expedition was a combined operation involving the fields of botany and geology. The great distances covered and the short time available for field work made it impossible to obtain a complete record at any one place. Nevertheless, the Highway offered an opportunity to make a rapid reconnaissance across northwestern North America—the northern plains, the mountains, and the subarctic interior lowlands—and to prepare a brief if incomplete account of the surficial geology. Highway construction was in progress, and many fresh roadcuts were available for study.

The geological objectives were a knowledge of the surficial geology along the Highway, and its interpretation to give a tentative late Cenozoic chronology that would serve as a basis for the botanical work. Another objective

was knowledge concerning past and present-day frost modification of the terrain.

The botanical objectives were, first, to gather the basic floristic data with which to describe the flora of this little-known region and to define the ranges of its species; second, to outline the geographic distribution of plant communities and to consider their historical development. Plant communities are not static but vary with changes in soils, topography, and climate.

Certain joint or "geobotanical" objectives of the expedition are especially significant. The development of the flora and its arrangement into plant communities are closely related to the origin and subsequent modification of the land surface and the soils. Movement of the soil by frost action is enormously significant in conditioning the development of the vegetation, and probably retarded it for a long period before the land was covered by modern forests. What now goes on above or beyond timber line is a clue to the nature of these past changes and their effect on vegetation. The extension of forests over tundra may cause changes in permafrost or in frost processes, and need not necessarily be the result of such changes.

The botanist from his study of floristic geography, of plant communities, and of the known or inferred behavior of species under varying environmental conditions, builds up a botanical chronology that must be reconciled with the geological chronology. The inferred presence of nunataks, for example, has often not been in accord with geological ideas about the location of unglaciated areas. The botanical evidence should be weighed carefully by the geologist because it may represent time intervals as yet unrecognized in a study of the glacial deposits.

The results of the botanical surveys by Raup and his assistants will be published in detail elsewhere. For the present report, however, Raup has been kind enough to prepare a short account of the vegetation.

An attempt was made to identify the various types of vegetation on aerial photographs and to correlate these types with the glacial deposits. This geobotanical study brought to light data useful in interpreting the various kinds of terrain shown on aerial photographs of this and similar regions (Raup and Denny, 1950).

In 1944 a second expedition, covering the section of the Alaska Highway between Whitehorse, Yukon, and Fairbanks, Alaska, was conducted by H. M. Raup. The geological work was carried on by J. H. Sticht (1951), then of the Department of Geology and Geography in Harvard University, and now at the College of the Pacific, Stockton, California. Frederick Johnson, archaeologist from the Robert S. Peabody Foundation for American Archaeology at Andover, Massachusetts carried on archeological work during 1944.

Field Work

Field work was confined to the country immediately adjacent to the Highway. Base camps were established at selected sites for periods of several days to nearly four weeks. Short stops were made elsewhere.

The party left Dawson Creek, British Columbia, on June 8 and camped on the Beatton River from June 9 until July 4. The period July 7–28 was spent at Summit Lake, about 90 miles west of Fort Nelson; July 30–August 7 at Watson Lake, Yukon Territory, in the upper valley of the Liard River; August 9–21 at a point about 50 miles southeast of Whitehorse between Teslin and Little Atlin lakes. About 2 weeks were spent on the return trip to Dawson Creek, where the party arrived on September 5. A mile-by-mile traverse was prepared to show the character of materials exposed in roadcuts. Aerial photographs of the region from Fort Nelson to Whitehorse were used as base maps on which to plot the glacial geology and the vegetation. In the field, vertical control was obtained with a Paulin altimeter from known elevations along the Highway. Where possible, these have been checked with the Alaska Highway Plans of the U. S. Public Roads Administration.

Personnel

The expedition was under the direction of Hugh M. Raup of Harvard University. Lucy G. Raup collected lichens and served as general assistant in the botanical work. The writer carried on the geological investigations. Donovan S. Correll, then of the Botanical Museum at Harvard, and now with the U. S.

Department of Agriculture, assisted in the botanical work. Karl and David Raup served as camp assistants.

Acknowledgments

The expedition was financed by generous assistance from several sources. The geological work was made possible through a grant from the Penrose Fund of The Geological Society of America. The botanical work was supported by grants from the Arnold Arboretum, the Milton Fund of Harvard University, the General Science Fund of the American Academy of Arts and Sciences, the American Philosophical Society, the Bache Fund of the National Academy of Science, and the Society of the Sigma Xi.

The expedition was under the combined auspices of the Arnold Aboretum, The Geological Society of America, and the Joint Economic Committees of Canada and the United States. Through the Joint Economic Committees, transportation by the United States Army over the Highway was procured, as well as authority to purchase supplies at commissaries. Splendid support was given by Wendell E. Thorne, Executive Secretary of the United States committee of the Joint Economic Committees. Special thanks are due the officers and men of the Northwest Service Command, United States Army, for their excellent cooperation and for the many courtesies extended to the party on the Highway. To the members of other field parties sent out by the Dominion and British Columbia Governments special thanks are also due: notably to C. O. Hage, C. S. Lord, and M. Y. Williams of the Geological Survey of Canada; to A. Leahey, Chief of the Soils Division of the Department of Agriculture, Canada; and to C. H. D. Clarke and H. L. Holman of the Lands, Parks, and Forests Branch of the Department of Mines and Resources. Camping equipment was loaned to the party by the Geological Survey of Canada. Aerial photographs were obtained from the U. S. Air Force through the assistance of the Joint Economic Committees and the Military Geology Branch of the U. S. Geological Survey. The author is most grateful to Wesleyan University for granting him a leave of absence to carry on the field work, and to the U. S. Geological Survey for the opportunity to complete the manuscript. Valuable information has been supplied by the U. S. Public Roads Administration through the kindness of R. E. Royall and W. T. Pryor. The manuscript has had the benefit of constructive criticism by H. S. Bostock, R. F. Black, and L. C. Peltier. The author is indebted to the late Kirk Bryan for his continued assistance throughout this study. Finally, to the other members of the expedition and especially to H. M. Raup, the author is indebted not only for help and inspiration in the field but for sound advice and assistance in the preparation of the manuscript.

Physical Geography

Topography

Alberta Plateau: Dawson Creek to Fort Nelson.—The interior plains of western Canada (Hume, 1944), between the Canadian Shield and the Rocky Mountains, have been called the Alberta Plateau in northeastern British Columbia, Alberta, and Saskatchewan (Camsell and Malcolm, 1919). The Alaska Highway from Dawson Creek to Fort Nelson is within the plateau province except where it follows a southward extension of the Mackenzie Lowlands (Hage, 1945) in the broad valley of the Muskwa and Prophet rivers. Along the Highway, the Alberta Plateau is underlain by gently east-dipping Cretaceous sandstone, shale, and pebble conglomerate[1], maturely dissected by such east-flowing streams as the Beatton, Sikanni Chief, Buckinghorse, and Prophet rivers (Pls. 1, 3). South of Fort St. John, the Peace River and its tributaries have carved broad lowlands, and only small plateau remnants remain.

Between the Alberta Plateau and the Rocky Mountain front there is a transitional belt characterized by long north-trending ridges and broad subsequent valleys. Along the Peace River, the eastern edge of the Rocky Mountains province lies near Hudson Hope, 40 to 50 miles west of Fort St. John. For about 40 miles west

[1] Unless otherwise noted, bedrock data given in this section for the region south of Fort Nelson are taken from Hage (1944) and for the region between Fort Nelson and Summit Lake from Williams (1944).

a belt of moderately folded Triassic and Cretaceous rocks form the Peace River Foothills (McLearn, 1940) and range in altitude from 1800 to 5000 feet. To the west the high peaks of the Rockies are predominantly sedimentary rocks of Paleozoic age.

Between Beatton River and Trutch Creek (Pl. 1), a steep, west-facing escarpment, as much as 1000 feet high, separates the Alberta Plateau from the lowlands to the west that have been carved in weak Lower Cretaceous shale and contain some elongated hills, such as Pink Mountain south of the Sikanni Chief River (Fig. 4). To the south this lowland is occupied by the Halfway River and its tributaries and to the north by the upper Prophet and Muskwa rivers. All these streams have cut headward into the weak Lower Cretaceous shale as subsequent tributaries to the main, eastward drainage from the Rocky Mountains. Between the Beatton and Prophet rivers, the eastern edge of the mountains lies only about 12 to 18 miles west of the Highway. Resistant Triassic rocks form peaks that are probably equivalent to the Peace River Foothills and have a relief of about 3000 feet. Hage (1944) includes these peaks in his mountain province.

Williams has divided the region west of Fort Nelson into an eastern, or plains province, the Foothills belt, and the Rocky Mountains section. The eastern plains, part of the Alberta Plateau province, extend more than 50 miles west from Fort Nelson and are "underlain by nearly flat-lying soft, sandy shales of Lower Cretaceous age. . . . The Foothills belt, about 50 miles wide, consisting of a deeply dissected Tertiary peneplain. . . . is underlain by little-disturbed Cretaceous shales and sandstones in the east and by faulted and folded Mesozoic and Paleozoic formations in the west." (Williams, 1944, p. 4). The Rocky Mountain section, beginning just east of Summit Lake, is underlain by a thick series of pre-Cambrian and Paleozoic sedimentary rocks.

Rocky Mountains: Summit and Muncho lakes.—The Highway between Fort Nelson and the Lower Crossing of the Liard River passes by way of Summit and Muncho lakes through the northern end of the Rocky Mountains which end a short distance south of the Liard River. Summit Lake is on the divide between the Tetsa River and MacDonald Creek, a tributary of the Racing River to the west. The divide just west of the lake has an altitude of about 4250 feet, the highest point along the entire Highway. The peaks northwest of Summit Lake reach altitudes of about 7000 feet, and Mount St. George, about 8 miles southwest of the lake, rises to nearly 8000 feet. Massive limestone and interbedded sandstone (Middle Silurian and Middle Devonian) form a broad anticline. The rocks between Summit Lake and MacDonald Creek are more or less flat lying. Just east of Summit Lake the rocks dip steeply eastward along the fault at the mountain front[2].

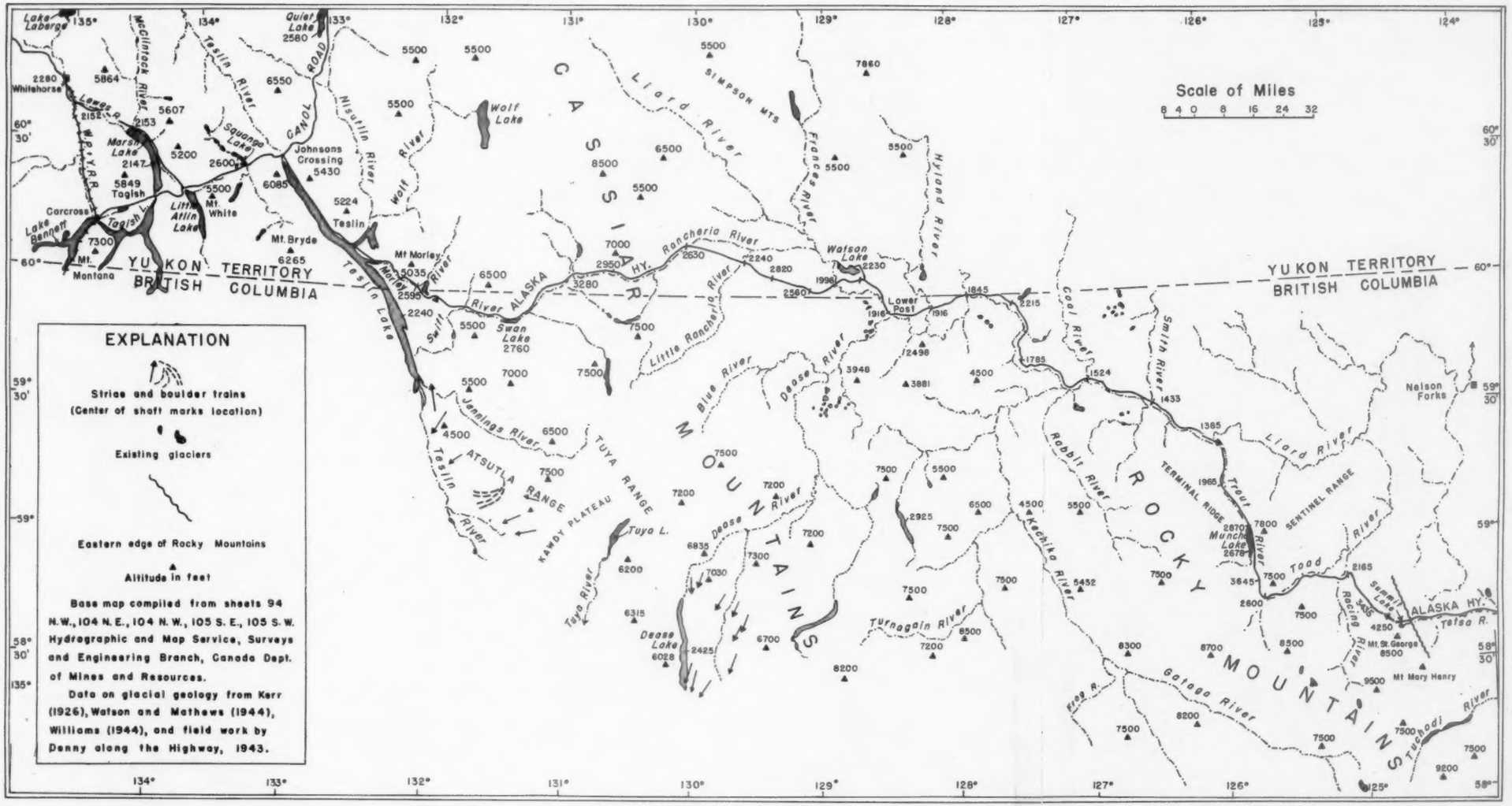
Westward from Summit Lake, the Highway descends steeply to the floor of MacDonald Creek[3] which it follows to the junction with the Racing River about 30 miles northwest of Summit Lake. The Highway leaves the valley of the Racing River and passes through a broad gap in the foothills to enter the valley of the Toad River. The Toad River-Racing River divide is a flat-floored, grassy meadow that appears to be an old river channel through which melt water flowed between the Toad and Racing River valleys. A few miles west of this divide, the Highway enters the canyon of the Toad River, which it follows for about 15 miles and then turns northward up a tributary canyon to the Toad River-Muncho Lake divide at an altitude of about 3645 feet.

Muncho Lake lies in a deep north-trending valley, about 14 miles long and 1 to 2 miles wide, that is bordered by mountains which rise to heights of 7500 to 8000 feet. East of the lake, the mountains are composed of massive Middle Silurian and Middle Devonian limestones; to the west, sandstones and conglomerates of Cambrian (?) age are present. On the east side of the lake and northward, the rocks strike approximately north and form steep dip slopes through which small westward-flowing streams have carved deep canyons. Small cirques are visible on the peaks west of the lake and west of the upper Trout River Valley to the north.

Great alluvial fans have been built on the east side of the valley where westward-flowing

[2] Unless otherwise noted, bedrock data given in this section are taken from Williams (1944).

[3] Called "McDonnell River" on Liard River sheet (94 N.W.), National Topographic Series, Canada Dept. Mines and Resources, 1944.



MAP OF AREA FROM SUMMIT LAKE, BRITISH COLUMBIA, TO WHITEHORSE, YUKON TERRITORY

streams leave the mountains (Fig. 5). These fans either reach to the eastern shore of the lake or come out on top of or cut through the terraces along the east side of the Trout River to the north. Some fans are as much as a mile in diameter at their lower ends along the lake shore and are partly overgrown by a spruce forest. However, many fans are almost completely devoid of vegetation and were apparently more or less covered by streams during the torrential rains of early July 1943.

Liard Valley: Lower Crossing to Watson Lake.—The Liard Valley marks the northern end of the Rocky Mountains and separates them from the Mackenzie Mountains to the north and northeast. It has long been used as an avenue of travel between the interior of Canada and the Yukon Basin. Johnson's (1946) archaeological finds suggest that such movements began in prehistoric times. The low country immediately bordering the river is fairly narrow near the Lower Crossing, but near Watson Lake it widens out into a broad plain that has been dissected to depths of several hundred feet by the Liard and its tributaries.

From the Lower Crossing of the Liard, the Highway follows the north bank of the river to Watson Lake, a distance of about 140 miles. Westward to Coal River (Pl. 2), the Highway follows a low terrace along the river, that here flows in a narrow valley bordered by mountains, some 2000 or 3000 feet high. The region is underlain by shales and sandstones of Upper Devonian and Mississippian age and by Middle Silurian and Middle Devonian limestone and related rocks. Hot springs are located on the north side of the Liard Valley near the Lower Crossing (Williams, 1944, p. 31–32).

West of Coal River, the Liard Valley is slightly wider. The bordering uplands are rolling and rise to heights of not more than a thousand feet above river. In places the Highway leaves the flood-plain and is built along the top of the surrounding terraces or on the higher till-covered upland where shale is exposed in road-cuts. The till is locally a rather heavy, silty clay. The upland supports a forest of black spruce, white spruce, and lodgepole pine, and there is a rather close coincidence between the presence of black spruce and the heavy till (Raup, 1945, p. 35). This relationship exists in scattered places along the Highway as far west as the Hyland River (Pl. 3).

West from Iron Creek the bordering highlands die out, and the Liard flows through a rolling upland of moderate relief that extends many miles north and south and is underlain by Upper Devonian and Mississippian shales and related rocks. For much of its length the Highway follows the edge of a broad, pitted outwash plain into which the Liard is incised to depths of 200 to 300 feet.

Watson Lake (Fig. 7) is near the center of the upper Liard watershed, that forms a gap between the Rocky and Cassiar mountains to the south and the Mackenzie Mountains to the northeast. This broad lowland is characterized by broad dissected outwash plains, great areas of which have been burned and are covered by a forest of lodgepole pine.

Cassiar Mountains: Rancheria-Swift-Morley River drainage.—From the Upper Crossing of the Liard, about 3 miles southwest of Watson Lake, the Highway passes through the Cassiar Mountains over the Rancheria-Swift-Morley River divide to enter the headwaters of the Yukon Basin near Teslin Lake (Pl. 2). The Cassiar Mountains have a granitic core[4] but near the Highway rise to only moderate heights. From the Upper Crossing the Highway traverses a rolling outwash plain and till-covered upland (Pl. 2) for about 40 miles to the valley of the Rancheria River, which it follows westward to the Continental Divide.

From the Lower Crossing of the Rancheria River westward, the valley, only a few miles wide, is bordered by relatively low rounded peaks, which are forest-covered nearly to their summits. The country rock is mainly limestone and related sedimentary rocks. About 10 miles west of the lower crossing, the valley narrows and the surrounding peaks increase in altitude toward the eastern edge of the granitic core of the Cassiar Mountains, where numerous small cirques indent the north slopes of the peaks to the south of the Highway.

An area of black olivine basalt is present on the floor of the Rancheria Valley in the granitic core of the Cassiar Mountains. Nowhere along the Rancheria or Swift rivers does the basalt occur more than 400 feet above the valley

[4] Unless otherwise noted bedrock data given in this section are taken from Lord (1944).

bottoms. "Glacial erratics of similar basalt are common up to elevations of about 5200 feet and the rock must, therefore, be older than at least a part of the Pleistocene" (Lord, 1944, p. 14).

Between the Rancheria and Swift rivers, the Continental Divide is asymmetric, the Rancheria draining eastward from the divide at a considerably higher level than does the southwest-flowing Swift River. The Swift River has proceeded by headward erosion almost to the headwaters of the Rancheria River; indeed, just prior to the last glaciation, piracy may have taken place.

The Swift River[5] flows in a narrow valley southwest from the divide to Swan Lake, which is drained through a rather broad valley that empties ultimately into Teslin Lake. The Highway leaves this valley near Morley Lake and follows the Morley River to the shore of Teslin Lake near Nisutlin Bay.

Teslin Mountains: Teslin Lake to Whitehorse.—In the Teslin Mountains (called "Teslin Plateau" by Bostock, 1948, Map 922A), that lie near the headwaters of the Yukon River system, the major drainage lines and the big lakes follow long, narrow valleys that trend north and northwest, apparently parallel to the major structure of the region (Cockfield and Bell, 1926). The upper Yukon Basin is characterized by valleys incised below an old erosion surface, the Yukon Plateau. Along the Highway, remnants of this plateau are conspicuous in only a few places.

Teslin Lake lies in a relatively straight, narrow valley that extends from the headwaters of the Stikine River in northern British Columbia to the Lewes River in southern Yukon Territory about 250 miles (Lees, 1936). The lake has an altitude of about 2240 feet, is 60 miles long, and drains northward through Teslin River to join the Lewes River about 65 miles north of Whitehorse. A few remnants of the Yukon Plateau lie about 2750 feet above the lake (Lees, 1936).

[5] Called "McNaughton River" on Dease Lake sheet (104 NE), National Topographic Series, Canada Dept. Mines and Resources, 1945. Swan Lake is called "McNaughton Lake" on this map. The mountains around Dease Lake are called the Cassiar Mountains on Plate 2, following the usage of the geological reports on the Dease Lake area. On the Dease Lake sheet they are labeled "Stikine Mountains."

The Highway follows the northeast side of the lake from Teslin Post Office to Johnson's Crossing where it turns west and southwest to follow a broad valley to Squanga Lake, and thence by a narrow U-shaped valley just north of Mount White to the north end of Little Atlin Lake. At the north end of Little Atlin Lake, the Highway turns northwest to the eastern shore of Marsh Lake which it follows to the place where the lake empties into the Lewes River.

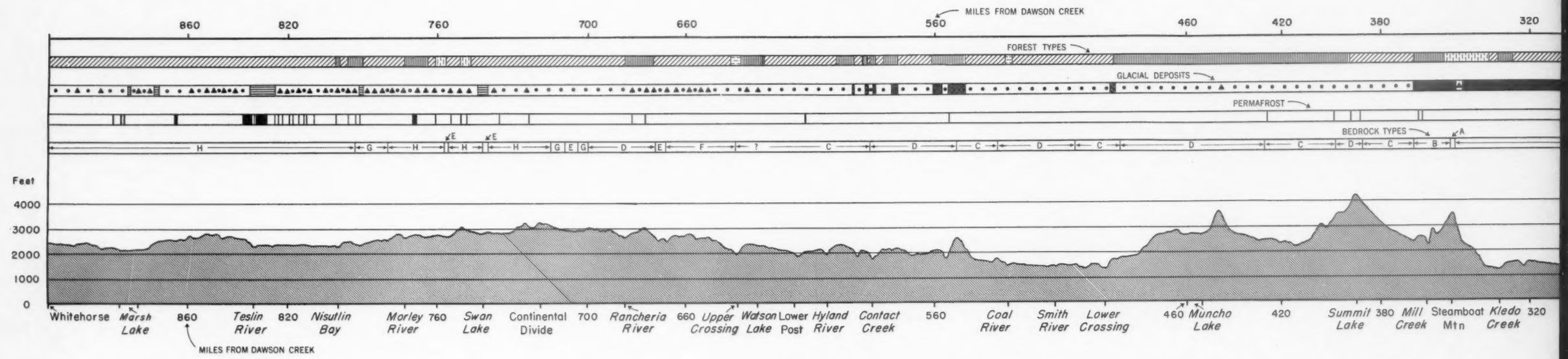
Marsh Lake, altitude about 2150 feet, is bordered by mountains that are 3000 feet or more above the lake. A valley now filled with pitted outwash deposits extends from the north end of Little Atlin Lake to the eastern side of Marsh Lake. The Lewes River, from Marsh Lake to Whitehorse, is bordered in most places by a wide terrace of glacial debris, indented by numerous kettle lakes. The Whitehorse area is surrounded by relatively low mountains with more or less rounded summits.

Climate

The climate of the region is subarctic, characterized by relatively short summers and long, dry winters (Table 1). Precipitation is low, 11 to 16 inches a year, and about a third falls as snow. In warmer climates such a light precipitation would result in semi-arid conditions, but in British Columbia and Yukon the low rate of evaporation and the perennially frozen subsoil are responsible for abundant surface water and swampy ground.

The low rainfall and rather high summer temperatures produce, however, several features more commonly associated with areas such as southwestern United States that have a semi-arid climate. Solution facets are abundant on limestone talus boulders near Summit Lake (Pl. 12, fig. 3). Solution-faceted limestone pebbles (Bryan, 1929) were found above timber line near Summit Lake and on alluvial fans and terraces north of Muncho Lake. Heavy rains in July 1943 caused severe floods on the eastern slopes of the Rockies in the Summit area. The vegetation indicates that these floods are infrequent.

Temperatures range from nearly 70° below zero to more than 100° above zero (Table 1, Fig. 2). Mean winter temperatures are lowest at lower altitudes, in valleys (Fig. 3). The



EXPLANATION

FOREST TYPES

White spruce—lodgepole pine—aspen

White spruce—balsam poplar (on flood plains)

Black spruce—white spruce—lodgepole pine

White spruce—alpine fir—black spruce—lodgepole pine

White spruce—balsam poplar—aspen

Aspen groves—prairie openings

GLACIAL DEPOSITS

Clay till or boulder clay. Up to 10% stones in clay and silt matrix. Sandstone pebbles predominate. Soils are generally only 2 to 3 feet thick and consist of loams, silts, and fine sands over clays.

Boulder till. Up to 30% stones in a sandy or silty matrix. Sandy, silty, and clayey soils generally not over 2 feet thick.

Varved sediments. Bedded fine sand, silt, and clay deposited in glacial lakes. Soils, 2 to 3 feet thick, vary from loams to heavy clays.

Gravel and sand on terraces and flood plains. Many small deposits not included. Soils vary from gravels and sands to silts and clay loams.

A — Sandstone and conglomerate; Upper Cretaceous. South of Muskwa River includes top of Lower Cretaceous.

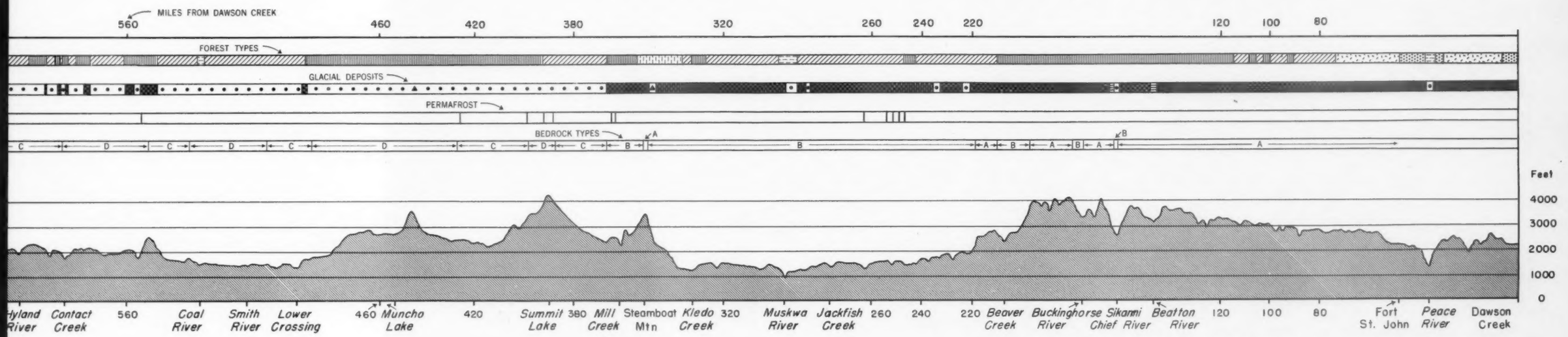
B — Shale, sandy shale and sandstone; Lower Cretaceous.

C — Shale, sandstone and limestone; Paleozoic and Mesozoic.

Limestone, sandstone

Thick glacial d

PROFILE OF ALASKA HIGHWAY BETWEEN DAWSON CREEK, BRITISH COLUMBIA, AND WHITEHORSE, YUKON TERRITORY



Profile drawn from Alaska Highway Plans, U. S. Public Roads Administration. Forest types after Raup (1945). Glacial deposits from road traverse by Denny and Alaska Highway Plans. Bedrock types generalized from maps by Hage (1944), Williams (1944), Lord (1944), Lees (1936), and Cockfield and Bell (1926). Areas of permafrost encountered during highway construction from U.S. Public Roads Administration.

EXPLANATION

GLACIAL DEPOSITS

Varved sediments. Bedded fine sand, silt, and clay deposited in glacial lakes. Soils, 2 to 3 feet thick, vary from loams to heavy clays.

Gravel and sand on terraces and flood plains. Many small deposits not included. Soils vary from gravels and sands to silts and clay loams.

BEDROCK TYPES

A — Sandstone and conglomerate; Upper Cretaceous. South of Muskwa River includes top of Lower Cretaceous.

B — Shale, sandy shale and sandstone; Lower Cretaceous.

C — Shale, sandstone and limestone; Paleozoic and Mesozoic.

D — Limestone, sandstone and conglomerate; chiefly Paleozoic.

E — Basalt; Tertiary.

F — Thick glacial drift; many areas not shown.

G — Granites; Jurassic and later.

H — Metamorphic and igneous rocks; pre-Cambrian and later. Small areas of limestone near Whitehorse and south of Mile 855–860. Large areas are drift-covered.

PERMAFROST

Area where permafrost was encountered during highway construction.

PROFILE OF ALASKA HIGHWAY BETWEEN DAWSON CREEK, BRITISH COLUMBIA, AND WHITEHORSE, YUKON TERRITORY

TABLE 1.—CLIMATOLOGICAL DATA*

Place	Temperature °F			Precipitation (Inches)					
				Summer Average (May-Sept)		Winter Average (Oct-April)			
	Mean Annual	Maximum	Minimum	Rain	Snow	Rain	Snow	Total (Water)	Mean Annual (Water)
Fairbanks, Alaska	26	99	−66	—	—	—	—	—	11.8
Whitehorse, Yukon	27	84	−69	7.09	3.5	0.50	34.6	3.96	11.4
Watson Lake, Yukon	31.4	90	−61	8.27	1.6	0.61	63.3	6.94	15.37
Fort Nelson, B.C.	30	98	−54	7.73	0.7	0.85	52.6	6.11	13.9
Fort St. John, B.C.	35	105	−65	9.86	1.4	0.86	48.6	5.72	15.7

* From Eager and Pryor (1945). Based on data obtained from Canadian Dept. of Transport and U. S. Weather Bureau for periods of 5 years or more.

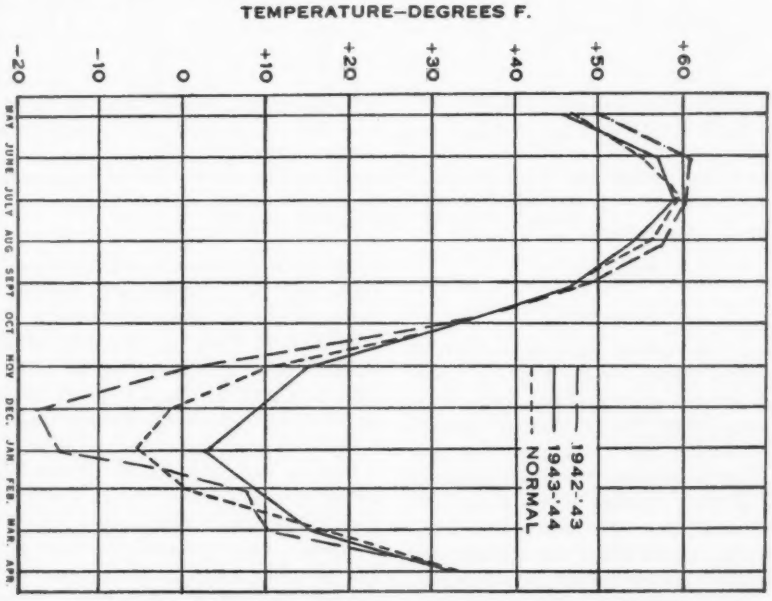


FIGURE 2. MEAN MONTHLY TEMPERATURE ON THE ALASKA HIGHWAY

Based on records at Fort St. John, Fort Nelson, Whitehorse, Northway, Tanacross, Big Delta, and Fairbanks (Eager and Pryor, 1945).

mean annual temperature is below 28° from Whitehorse northward and above 28° to the southeast. This is significant because permafrost (or perennially frozen ground) is widespread west of Whitehorse and present in only a few places to the south (Fig. 8 and Pl. 3).

In midwinter, approximately at latitude 60° N. (Watson Lake to Whitehorse), the sun is above the horizon for only 6 hours daily whereas in midsummer the sun is visible for 19 hours. The generally low altitude of the sun throughout the year accentuates the difference in insolation between north- and south-facing slopes. The latter commonly show more evidences of mass movement; the vegetation is not the same as that on north-facing slopes and

generally appears to be more disturbed by present-day frost heaving.

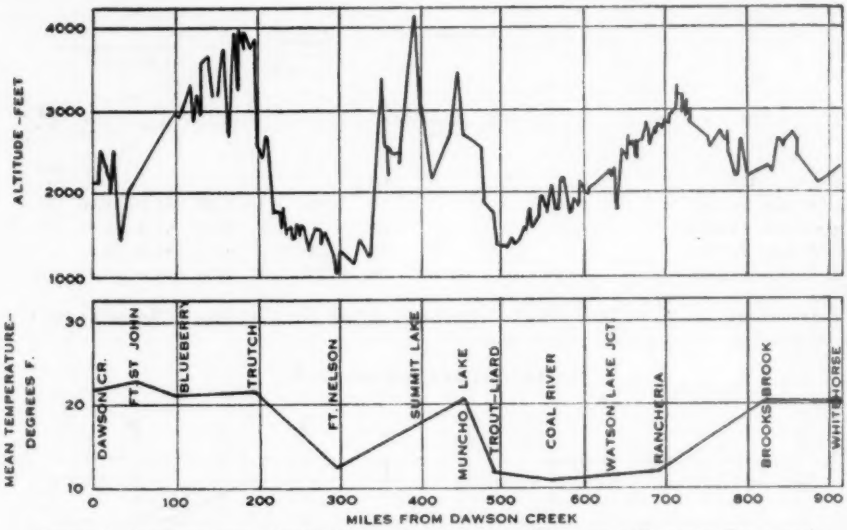


FIGURE 3. ALTITUDE AND MEAN WINTER TEMPERATURE (NOVEMBER THROUGH FEBRUARY) BETWEEN DAWSON CREEK AND WHITEHORSE
Winter of 1943–1944 (Eager and Pryor, 1945)

Vegetation

BY H. M. RAUP

Forests.—Throughout nearly all of its extent southeast of Whitehorse the Highway passes through forested country (Raup, 1945; Raup and Denny, 1950). The principal exception is at the southern end where there are agricultural prairie openings. The actual number of species of trees involved in these extensive forests is small. Only eight achieve abundance: white spruce (*Picea glauca* var. *albertina* and var. *Porsildii*), black spruce (*Picea mariana*), alpine fir (*Abies lasiocarpa*), balsam poplar (*Populus tacamahacca*), paper birch (*Betula papyrifera* var. *commutata* and var. *humilis*), trembling aspen (*Populus tremuloides*), lodgepole pine (*Pinus contorta* var. *latifolia*), and the northern larch (*Larix laricina* var. *alaskana*). Shrubs are, of course, very abundant, particularly the dwarf birch (*Betula glandulosa*), two species of alder (*Alnus crispa, A. tenuifolia*), and many species of willow (*Salix*).

The most abundant and widespread forest community between Dawson Creek and Whitehorse is of white spruce, lodgepole pine, and aspen (Pl. 3). It is best developed on better-drained soils at low and middle altitudes, at or below 2800 feet. It produces its heaviest and commercially most promising timber on higher river terraces along the Prophet, Muskwa, Nelson, and Liard rivers, at or below an altitude of 2000 feet. It is also well-developed on the divide between the Halfway and Blueberry rivers northwest of Fort St. John.

Next in importance, as far as area is concerned, is an association of black spruce, white spruce, pine, and in places aspen. This forest is of wide occurrence along the Highway but has its most extensive development in two districts, the first of which begins about 50 miles north of Fort St. John and extends over the Alberta Plateau north for about 150 miles. The other large area is in the rugged mountainous country between Summit Pass and the lower crossing of the Liard River. Smaller, scattered areas appear about 40 miles west of Fort Nelson, between Steamboat Mountain and the Tetsa River, between Coal and Hyland rivers, and between Watson Lake and Teslin. Nowhere does the black spruce-white spruce-lodgepole pine type produce forests of much

value. Most of its trees are small in stature and in many places widely spaced.

The forests of the high plateaus between Fort St. John and Fort Nelson are notable in ways other than the predominance of black spruce. Aspen is relatively unimportant and is abundant only on warm south-facing slopes. On the higher parts of the terrain, particularly above 3000 feet, paper birches are nonexistent. The valleys are characterized by rather broad plains that have an open timber of black spruce and lodgepole pine, interspersed with shrub communities of dwarf birch and willows.

Throughout nearly all of the country between Summit Lake and the Liard River, the topography is extremely rugged and mountainous. The only forests of any size are in the immediate valleys of the streams and are composed of white spruce and balsam poplar. The forests of the steep mountain slopes are nearly all of small stature and are various mixtures of white and black spruce, lodgepole pine, and aspen, the last chiefly on warm south-facing slopes.

Slope forests of excellent quality were seen in the Steamboat Mountain area 40 to 50 miles west of Fort Nelson, and some of lesser extent on the divide between the Morley and Swift rivers. All these forests are on foothill slopes below 3500 feet. They are composed principally of white spruce, fir, black spruce, lodgepole pine, and paper birch, the first two of which are the important timber-producing species. Trees more than 100 feet tall, with breast-height diameters of 14 to 20 inches, are common.

Muskeg or bog forest everywhere is characterized by black spruce, with or without an admixture of larch.

Another variation in forest types occurs in the Peace River agricultural districts where there are a few prairie openings, an abundance of balsam poplar in upland forests, and an extensive development of aspen on burned-over land and at prairie margins. The only suggestion of this type north of the Fort St. John district is on the till-covered upland just west of Fort Nelson, where there are heavy aspen forests and balsam poplar is occasional.

The only timber of commercial value is on river flood plains and on some foothill slopes. On the flood plains of the Prophet, Muskwa, Liard, and Rancheria rivers, the white spruces grow to heights of 100 feet or more, with breast-height diameters of 18 to 24 inches. On the foothills, the alpine firs and spruces attain similar dimensions and are of good timber quality. Most of the upland timber along the road, however, is much smaller and will produce only small and scattered quantities of good lumber.

Detailed examination of the widespread mixtures of lodgepole pine, aspen, and the spruces yields two outstanding facts relating to their history. First, in most of the stands where pine and aspen occur and have grown to maturity, they are not reproducing themselves. Apparently the aspen does so only in some of the semi-open prairie country of the Peace River district, and the pine only on the driest sandy or gravelly outwash plains and morainic ridges. Second, nearly all forests traversed by the road have been burned at one time or another. Observation of new and recent burns shows that pine and aspen owe their great abundance to the incidence of fire. In most cases they represent an initial stage of forest development that passes through only a single generation after fire. Their actual abundance in a particular stand of timber is commonly determined by the varying intensities of the fires that have given them prominence.

Edible berries abound in the forest in season. The most important are the blueberry (*Vaccinium uliginosum* L.), mountain cranberry (*Vaccinium Vitis-Idaea* var. *minus*), mooseberry (*Viburnum edule*), crowberry (*Empetrum nigrum*), black currant (*Ribes hudsonianum*), red currant (*Ribes triste*), rasberry (*Rubus strigosus*), gooseberry (*Ribes oxyacanthoides*), and strawberry (*Fragaria glauca*).

Prairies and alpine tundra.—Although the immediate borders of the Highway are nearly all forested, semi-open prairies or arctic-alpine tundra may be easily reached at many points.

Most of the agricultural development of the Peace River region is on former natural prairie, remnants of which are still to be seen along the road between Dawson Creek and Fort St. John. Their principal species are drawn from the "mid-grasses," such as *Agropyron*, *Stipa*, *Koeleria*, and *Calamagrostis*. A few grassy openings are seen on heavy lacustrine soils between Fort St. John and Fort Nelson. Most of these

seem to be due to fire but some along the Beatton River appear to be natural.

Timber line at Summit Pass is only about 300 feet above the road. Above timber, the vegetation is either an alpine scrub of dwarf birch and willow or an alpine tundra of mosses, lichens, grasses, and sedges. There are large areas of bare rock and talus, but even these contain small patches of alpine tundra. Various arctic soil forms, such as stone rings, stone stripes, and related features are characteristic of the surface and are intimately related to the development and distribution of the vegetative cover.

The Continental Divide, separating the waters of the Rancheria from those of Swift River, is at an altitude of about 3000 feet and contains extensive muskegs. Most of the higher mountain slopes are some distance from the road, and not easily accessible. However, some low summits near the head of the Swift River can be reached without much difficulty, and at one place between Teslin and Little Atlin lakes, there are easily accessible alpine areas on either side of the road.

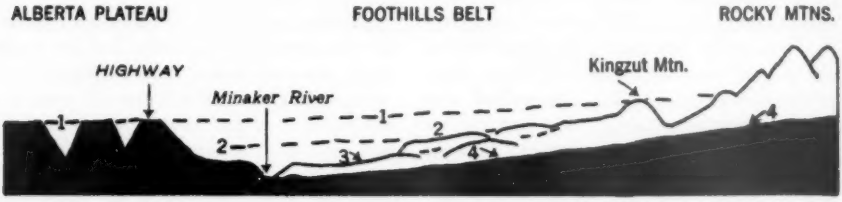


FIGURE 4. DIAGRAMMATIC PROFILE OF EROSION SURFACES IN THE VICINITY OF KINGZUT MOUNTAIN, BRITISH COLUMBIA

PRE-GLACIAL PHYSIOGRAPHIC HISTORY

Alberta Plateau: Dawson Creek to Fort Nelson

Widespread erosion surfaces of low relief, now dissected to form plateau remnants, are characteristic of the region adjacent to the Highway from Dawson Creek to Whitehorse. East of the Rocky Mountains, the oldest surface is the Alberta Plateau (Pl. 1), a maturely dissected plateau whose surface has a slight eastward slope that was cut prior to the excavation of the lowlands of the west (No. 1, Fig. 4). Williams (1944), by analogy with similar surfaces farther south, believes it to be of Miocene age. No pre-glacial gravel deposits were found on the plateau, except for one locality reported by Hage[6] about 6 miles north of the Sikanni Bridge.

Steamboat Mountain and adjacent peaks west of Fort Nelson are thought by Williams to be remnants of an old land surface, perhaps related to the plateau remnants between the Muskwa and Prophet Rivers and northwest of Fort Nelson. According to Williams (1944, p. 10), this "plains surface" extends westward across the foothills, and just east of the Rocky Mountain front has an altitude of about 5,400 feet. It "... is tangent to the upper rounded mountain tops of the Rocky Mountains." It is the writer's impression that Williams' "plains surface" crosses the foothills belt and passes into the mountains about 1000 feet below their summit (No. 1, Fig. 4), and is the same surface as that found southwest of MacDonald Creek about 10 miles west of Summit Lake.

The foothills belt (Fig. 4) is a subsequent lowland dissected by north-south tributaries of the major east-flowing streams. At least two east-sloping erosion surfaces (Nos. 2 and 3) were carved during the development of the foothills belt. The higher level (No. 2) in the vicinity of Kingzut Mountain is probably the equivalent of the floor of the Beatton Valley and the broad valley of the Sikanni Chief River above the canyon. This old "broad valley" surface is now being dissected by tributaries of the Halfway River and by the Sikanni Chief.

[6] In a small pit on the south side of the Highway, "gravels composed of sandstone, quartzite, chert, limestone, and vuggy quartz are cemented by iron carbonate and underlie boulder till" (C. O. Hage, written communication, November 8, 1943).

It lies nearly 1000 feet below the top of the Alberta Plateau to the east and corresponds apparently with the outcrop of the Sikanni formation (sandstone, shale, and thin pebble conglomerate; Pl. 1). The lower surface (No. 3) is crossed by the Highway west of Steamboat Mountain along Mill Creek and the Tetsa River.

A wind gap across the north end of Pink Mountain west of the Beatton River (Pl. 1) suggests that the Halfway River has committed piracy on the headwaters of the Beatton. Similar drainage changes have pirated the headwaters of Trutch Creek, Buckinghorse River, and other east-flowing streams.

Uplift of the mountains caused deep dissection in the foothills belt at least prior to the last stage of mountain glaciation. Williams (1944, p. 12) considers the last uplift to be late Pliocene.

Rocky Mountains—Liard Valley—Cassiar and Teslin Mountains

Remnants of an old upland or dissected plateau surface are extensive in the Rocky Mountains southwest of MacDonald Creek. However, from the junction of the latter with the Toad River to Watson Lake, remnants of upland surfaces are not conspicuous near the highway (Pl. 2). Williams (1944) holds that the broad valley of the upper Toad River, south of the highway crossing, was once continuous with the Muncho Lake-Trout River Valley, and that, in pre-glacial time, piracy was committed on the headwaters of the Toad by the Racing River, into which the Toad River now flows by way of a deep narrow canyon.

South of the Rancheria, east of the Continental Divide, remnants of an old upland surface bevel the upper mountain slopes at altitudes of about 4500 feet (Lord, 1944). South of Swan Lake, remnants of an old upland surface, high on the slopes of the mountains, rise nearly to the floors of the cirques that indent the peaks that project above the uplands.

Plateau remnants are widespread in southern Yukon and adjacent parts of northern British Columbia between the Cassiar Mountains to the east and the Coast Ranges to the west (Kerr, 1936; Cairnes, 1912; 1913; Watson and Mathews, 1944). Near Teslin Lake, remnants of the Yukon plateau lie at altitudes of about 5000 feet, or about 2750 feet above the lake (Lees, 1936). Mount White, east of Little Atlin Lake, has an almost level summit plateau, which stands at an altitude of about 5500 feet. This remnant is several miles across and is underlain by limestone (Cockfield and Bell, 1926). On the upper slopes of Mount Montana south of Carcross, a remnant of the plateau rises into the cirques on the sides of the peaks that surmount the upland.

In summary, in later Tertiary time the region from Dawson Creek to Whitehorse was one of moderate relief with very broad lowlands. Subsequent uplift, probably for the most part pre-glacial, was more pronounced in the Rocky Mountains to the south near the Peace River than to the north near the Liard River (Hedley and Holland, 1941). Watson and Mathews (1944) suggest that the Teslin Trench has been dropped down relative to the surrounding plateau remnants in late Tertiary or possibly early Pleistocene time. The surfaces east of the Rocky Mountains suggest several stages of uplift, the last possibly Pleistocene in age.

Glaciation

Alberta Plateau: Dawson Creek to Fort Nelson

Summary.—The drift is dominantly a slightly weathered, bouldery clay till deposited by ice that came from the east. No extensive moraines or outwash deposits were seen near the Highway. The western limit of ice-sheet drift lies near the eastern edge of the Rocky Mountains. Hage's observations (1944) suggest that the last eastward advance of Rocky Mountain glaciers to the mountain front preceded the maximum westward advance of the ice sheets in Wisconsin (?) time, and that these two advances were separated by an interval of time when glacial lakes were formed in the foothills belt. Much of the drift may be pre-Altamont in age and probably was subjected to vigorous frost action during the Altamont substage when the area was surrounded by ice to the north, east, and west. Frost action and permafrost decreased in intensity and extent at some time before the present, perhaps with the advent of the post-glacial optimum and possibly coincident with

the development of prairies. A recent change in vegetation from prairie to forest (Raup, 1946) suggests a slight cooling of the climate in recent time.

Till.—The till is dominantly a boulder clay that contains from 3 to 10 per cent of sandstone pebbles. Unweathered till is a compact blue-gray boulder clay containing striated and somewhat rounded fragments of rock. The till is oxidized to depths of 5 to 8 feet and leached of calcium carbonate to depths of 2 to 4 feet. Most rock fragments in the weathered zone are fairly fresh, and some have a thin weathered rind. Striated stones are abundant. A faint cleavage or platty structure, more or less parallel to the ground surface, is visible near the top of some sections.

The maximum thickness of till exposed was 40 feet, along the Muskwa River. Hage (1944, p. 13) reports thicknesses of several hundred feet along the Prophet River. Wells drilled at the Fort Nelson airport (Hage, 1944) show that the drift there is about 700 feet thick. On top of the Alberta Plateau the till is thin but apparently rather continuous. Hage (personal communication) states that there is very little if any drift on the plateaus south of Trutch Creek east of the Highway and east of Indian Creek. Rock outcrops are few except along the principal streams.

The approximate contact between ice-sheet drift and Rocky Mountains drift is shown on the Glacial Map of North America (Flint *et al.*, 1945) as extending northwest as far as Hudson Hope on the Peace River (Pl. 1). Pebbles and boulders of granite, gneiss, and schist are fairly abundant in till along the Highway for about 80 miles north of the Peace River and in the area from the Buckinghorse River north to Fort Nelson and west almost to the eastern edge of the Rocky Mountains. In the Sikanni-Beatton region, however, pebbles of granite and gneiss are rare. Granite and gneiss boulders are especially abundant in the Prophet River Valley, in both till and gravel. West of the longitude of Pink Mountain along the Sikanni Chief River and to the south, Hage (1944) found no granite or gneiss pebbles in either till or gravel, although they do occur on the upper Buckinghorse River. West of Fort Nelson, Williams (1944, p. 26) found red granite and gneissic boulders in drift along the Tetsa River as far west as approximately 5 miles east of the Rocky Mountain front (7 miles east of Summit Lake). On Teepee Mountain they occur above an altitude of 4000 feet. Assuming that these granite, gneiss, and schist boulders are from the Canadian Shield to the east, the scattered observations suggest that the contact between ice-sheet drift and Rocky Mountains drift, or more accurately the western limit of ice-sheet drift (Keewatin or Laurentide; Flint, 1943), extends northwest from the vicinity of Hudson Hope to a point just east of Summit Lake, more or less following the eastern edge of the Rocky Mountains or the higher peaks of the foothills belt.

In the foothills belt west of the Highway, Hage (1944, p. 13) found two tills separated by lake deposits. The lower till, containing no granite or gneiss pebbles, is of Rocky Mountain derivation; the younger till, containing pebbles of granite and gneiss, was deposited by an ice sheet advancing from the east.

Lake deposits.—Varved sediments, consisting of stratified fine sand, silt, and clay, are exposed along the Highway in the valleys of the Beatton, Sikanni Chief, and Prophet rivers. West of the Highway Hage (1944) found bedded glacial lake clays in many places such as along the Halfway River about 3 miles east of Pink Mountain, along the Minaker River east of Kingzut Mountain, along the upper Sikanni River, and for 100 miles south of Fort Nelson in the valley of the Prophet River. All these deposits west of the Highway rest on till or gravel that contains pebbles of probable western (Rocky Mountain) derivation.

The flat floor of the Beatton Valley (Fig. 5; see also Raup and Denny, 1950, Pls. 14, 16) is underlain by brownish-gray, sticky clay with a granular or crumb structure that grades downward at depths of 5 to 8 feet into undisturbed, gray varved silt and clay (Table 2). In any one section the maximum number of varves visible or readily inferred is 100. At least 300 may be present in some places.

Gray clay is exposed in the Sikanni Chief Valley along the Highway about 2 miles north of the bridge. The exposures were badly slumped; no stratification was observed. The clay is slightly brownish at the top and forms the surface of the Sikanni lowland at elevations of from 400 to 500 feet above the river.

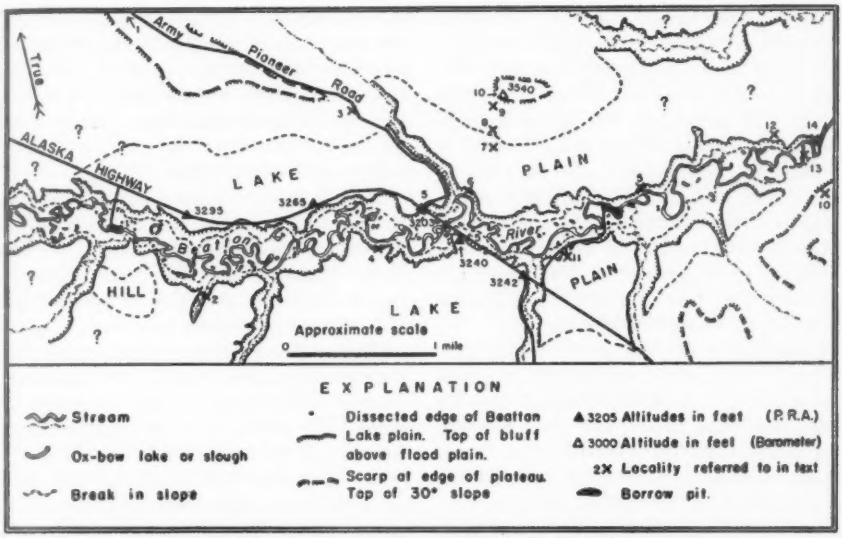


FIGURE 5. MAP OF BEATTON RIVER AREA, BRITISH COLUMBIA

Base compiled from aerial photographs taken by the U. S. Army Air Force, 26 May, 1944 (16DPU, 4M700, 5V145-5V173); and Alaska Highway Plans, U. S. Public Roads Administration. Location and scale approximate.

Localities referred to in text: (1) Fossil stone rings along army pioneer road about 1 mile to north. (2) Stone rings on bog iron deposit near spring. (3) Loose bouldery material (solifluction debris?) overlying stony clay till. (4) Slumps on undercut till bank. (5) Varved sediments. (6) Slumps in varved sediments. Ice lenses at depth of 4 feet on June 13, 1943. (7) Sandy silt in low knoll. (8) Ground frozen beneath 1 foot of moss, June 13, 1943. (9) Loose, sandy silt containing large sandstone slabs. Slope 30 per cent. Numerous steps or breaks. (10) Clay till, very few stones. (11) Gray, stony till, leached and iron-stained; about 10 per cent sandstone pebbles. (12) Brownish, silty sand in knoll. (13) 15 feet of gray silt and sand, faint stratification; overlying 5 feet of brownish, compact, stony till, deeply iron-stained. Pebbles of sandstone and one of schist and of granite. (14) 10 feet of gray silt and sand on gray clayey till.

The lake deposits that underlie the Beatton Valley were not observed in contact with till. However, it is assumed that they rest on the till exposed under the level of the lake plain at two localities (Fig. 5) where two pebbles of granite were found in the till. The smooth, nearly level surface of the Beatton Valley suggests that the valley floor is more or less an original depositional surface and has not been over-ridden by an ice sheet in post-lake time. Therefore the lake beds of the Beatton Valley are younger than a till deposited by ice from the east (Hage's younger till) and younger than the inter-till lake beds west of the Highway. The granular structure and absence of stratification in the upper part of the lake sediments along the Beatton River may be due to the disturbance of the silts and clays by freezing and thawing in post-lake time.

Terraces.—On the north bank of the Peace

TABLE 2.—VARVED SEDIMENTS EXPOSED IN ROAD CUT ON SOUTH BANK OF BEATTON RIVER, B. C. Locality on west side of Highway. Altitude—3250 feet

1 to 2 ft.	Stripped
1 ft.	Silty clay, no stratification, granular or crumb structure; roots
	-gradation-
6 ft.	Silt and clay, faint stratification, irregular and contorted; few sandstone pebbles
	-gradation-
4 ft.	Varves, about 1 inch thick consisting of a layer of silt and clay of about equal thickness; contemporaneous deformation locally; very delicate irregular laminations in the silt; a few small pebbles of sandstone up to half an inch in diameter

River, a gravel terrace about 200 feet above high water is underlain by well-sorted, bouldery and pebbly gravel and sand. A second, very narrow terrace is about 170 feet above high water. The stones are subangular to well-rounded in shape; some are as much as 2 feet in diameter. Oval-shaped stones also are present. A faint horizontal stratification is visible. Gray and yellow sandstone and quartzite are the predominant rock types together with gray porphyry, a pink granitic rock, black chert, and pink quartzite. Less extensive gravel deposits are found on the south side of the river.

Clean and well-sorted gravel consisting of somewhat rounded to well-rounded sandstone and quartzite pebbles and boulders caps the south side of the Sikanni canyon at an elevation of about 300 feet above the river. Similar gravel is also found south of the canyon up to altitudes of more than 500 feet above the river. Cross bedding and channeling are visible. Some pebble bands show an imbricate structure indicating an easterly stream flow. On the north side of the Sikanni Chief, only about 100 to 150 feet above river, are cross-bedded pebbly gravel and coarse sand, not as coarse-grained or as well-sorted as the deposits south of the river. The low-lying gravels are probably younger than those on top of the canyon, but their relationship to the clays about 2 miles to the north is uncertain. The gravel south of the river is found at an altitude at least as high as that of the clay. Hage's observations suggest that some of these quartzite gravels may be older than the lake deposits.

There are numerous outcrops of gravel and sand along the Highway from Fort Nelson southward almost to Trutch Creek. Some of the gravel contains numerous boulders of granite and gneiss. Hage (1944) found gravel beds resting on bedrock along the upper Prophet and the Minaker. These gravels contain pebbles of quartzite, sandstone, limestone, chert, and porphyry and are overlain by bedded glacial lake clays.

Age of drift.—At the present time it is not possible to correlate the drift in the Dawson Creek-to-Fort Nelson region with the several belts of end (terminal) moraine in central Alberta, shown on the Glacial Map of North America (Flint *et al.*, 1945). These morainic belts were mapped by Bretz (1943), who holds that the Altamont moraine in eastern Alberta and western Saskatchewan is the late Wisconsin terminal moraine. Antevs (1945) believes that all the youngest drift in the Dawson Creek-to-Fort Nelson region is post-Altamont; the slight amount of weathering of the drift is compatible with a late Wisconsin date.

There is slight evidence that the drift may be older than Altamont, that the time equivalents of the latter should be found to the east of the Highway, and that during Altamont time the Dawson Creek-to-Fort Nelson region was subjected to vigorous frost weathering and solifluction. The evidence for this hypothesis is threefold. First, the absence of any topographic expression for the till is striking. The till on the slopes of the plateau (as along the Beatton and Sikanni valleys) is overlain by several feet of loose, rubbly material that contains large sandstone slabs and is probably solifluction debris (Denny, 1938). This fact suggests some modification of these slopes by mass movements (solifluction) in post-till time, and the presence of a forest cover is a strong indication that such movement is not of modern date. Second, fossil boulder rings on top of the plateau north of the Beatton Valley are now almost completely overgrown by a forest of black spruce (Pl. 19, fig. 2), which indicates that they are relics of a past climate. Third, the granular structure and absence of stratification in the upper 5–8 feet of the lake sediments in the Beatton Valley suggests disturbance by frost action in post-lake time. In contrast, on the east side of Teslin Lake between Brook's Brook and Johnson's Crossing the lake sediments are well-stratified within a foot of the surface. Inspection of the Glacial Map of North America (Flint *et al.*, 1945) shows that the Altamont moraine or its time equivalents might extend from Edmonton, Alberta, in a northwesterly direction and pass to the east of Fort Nelson, perhaps to join the western glaciers somewhere in the north. Under this hypothesis the till along the Highway would be pre-Altamont. The post-till interval of wide spread permafrost and solifluction could then represent the Altamont substage when the Dawson Creek to Fort Nelson region was ice-free, but surrounded by glaciers on three sides (see also Raup, 1946).

The thick drift in the Prophet Valley is probably slightly younger than the till on the

plateau to the east, as it seems logical that tongues of the Laurentide ice sheet advanced and retreated up the main valleys across the plateau. Perhaps this valley drift is Altamont, and the plateau between the valleys was not covered by ice at this stage. This possibility finds support in Hage's statement (personal communication), that there is very little if any drift on the plateau south of Trutch Creek and east of Indian Creek. On the plateau between the Beatton and Sikanni Chief rivers, no erratics of eastern derivation were found.

Rocky Mountains: Summit and Muncho Lakes

Summary.—It is not known whether ice ever completely covered the higher peaks in the northern Rocky Mountains (Bostock, 1948, p. 10). In the Summit Lake area, erratics were found as high as about 5800 feet. Hedley and Holland (1941) believe that the northern end of the Rocky Mountains east of the Kechika River (Pl. 2) was completely covered by ice. In the Summit Lake area, striae indicate a general ice movement northward and eastward parallel to the main drainage lines. Williams (1944) reports that erratics from the Rocky Mountains are found many miles east of the mountain front. Striae suggesting an eastward movement were found by Williams near the Lower Crossing of the Liard.

The "old" till near Summit Lake is probably early Wisconsin. The deposits occupy the Tetsa River-MacDonald Creek divide (Pl. 4) where they are more or less protected from stream erosion. Their cementation by calcium carbonate has produced a rock that is resistant to weathering under the semiarid climate of the region. The surface of the "old" till does not show the constructional topography so characteristic of the deposits of the later glacial substages. Doubtless the gently rolling surface of the upland is the result of intensive frost action and solifluction that removed the original constructional topography during later substages of mountain glaciation when the surrounding valleys were occupied by glaciers.

Three retreatal substages during the last stage of mountain glaciation are suggested by the features in the Summit area. During the earliest substage, the ice extended far down the U-shaped valleys of MacDonald Creek and the Tetsa River. Only a few scattered patches of moraine have been recognized in the Tetsa Valley east of a point about 2 miles east of Summit Lake (Mile 387, see Pl. 4), such as at Miles 385 and 382 (Williams, 1944, p. 26). Eastward at least to Mile 370, the Tetsa River flows in a deep U-shaped valley with truncated spurs, probably the result of erosion by a valley glacier. The second substage is represented by a moraine on the north side of the valley about 2 miles east of Summit Lake (Pl. 4), at altitudes of 4400 to 4470 feet, or about 500 feet above the Tetsa River. High outwash terraces south of the Tetsa may have been formed at this time by melt-water that flowed east from MacDonald Creek valley south of the "old" till upland, across the valley of the North Fork Tetsa River, and through a gap in the mountains to the east (east of Loc. 4, Pl. 4). The third and most recent substage of the mountain glaciers is represented at an altitude of about 5800 feet by a small moraine that lies in the small cirque 2 to 3 miles southwest of Summit Lake. On the northwest side of this peak, a cirque is partly blocked by a small moraine at an altitude of about 5200 feet (Williams, 1944).

Two sets of terraces are present in the Tetsa Valley east of Summit Lake. The higher set is pitted; the lower set is not pitted and shows a channeled surface suggesting that it was formed during down cutting by the Tetsa River.

At some time during deglaciation, the Muncho Lake basin contained a valley glacier that extended northward down the Trout Valley at least to the Liard River. Terraces along the lower Trout River (Fig. 6) were built when there was ice in the immediate vicinity and perhaps indicate a stand of the ice front near Mile 477. The northern end of Muncho Lake apparently marks another position of the ice front when the gravel, sand, and silt that now underlie the terraces along the upper Trout River were deposited. Thereafter the ice in the Muncho Lake basin disappeared apparently without leaving any other large morainic deposits.

Summit Lake.—A boulder till (the "old" till) forms a rolling upland southwest of Summit Lake and underlies slopes on the north and south sides of the pass west of the lake (Pl. 4). Angular and subangular pebbles, cobbles, and boulders of limestone, sandstone, and quartzite,

some of which are striated and as much as 5 feet in diameter, occur in a small proportion of sandy matrix (Pl. 7, fig. 1). A few small sand lenses are present. The deposits are firmly cemented by calcium carbonate and have been eroded to form pillars or "hoodoos". The long, gently curving slopes (Pl. 6, figs. 1, 2) and deeply gullied edges of the till-covered upland southwest of the lake are distinct topographically from the steep-sided moraines and accentuated knob-and-kettle topography found elsewhere. The lime-cemented boulder till is overlain by moraines of a loose pebbly till that was deposited in valleys cut into the boulder or "old" till. The rather complete cementation of the "old" till probably took place when the latter filled the pass to a level several hundred feet above the lake. Ground-water conditions were apparently more favorable for cementation then than now.

Steep-sided moraines on slopes on either side of Summit Pass (Pl. 4) are underlain by a pebbly or sandy till that contains only a small amount of silt and clay. The pebbles of limestone and quartzitic sandstone are angular and subangular; a few have a "glaciated" form and are striated. A few of the sandstone pebbles are well rounded. Pits dug in these moraines commonly show a fairly well developed podzolic soil profile, with an inch of partly decayed vegetable matter underlain by a gray A-2 horizon about an inch thick, which grades downward into a brownish pebbly till. Where limestone pebbles are present the till matrix effervesces freely with dilute hydrochloric acid at a depth of about 1 foot below the surface. North of the pass the glacial deposits have a rather definite upper limit at altitudes of 4700 to 5100 feet.

The numerous steep-sided morainic ridges, shown on Plate 4 (Pl. 5, fig. 2), were formed by ice that rested in Summit Pass and was probably an offshoot of a glacier in the MacDonald Creek Valley. South of the pass, these moraines dam three small ponds in a valley eroded in the "old" till. Other moraines were formed by ice that came down the North Fork Tetsa Valley and descended eastward down the main valley of the Tetsa. Morainic ridges are notably absent from the surface of the "old" till upland.

At an altitude of about 5800 feet is a small moraine in the cirque on the north side of the peak 2 or 3 miles southwest of the lake (Pl. 4). On the headwall of this cirque, a triangular-shaped talus deposit extends out a little distance from the headwall. Boulders appear to have slid down over snow or ice. Possibly this is a protalus rampart, but it is probably too small to represent the last substage of a glacier in the small cirque.

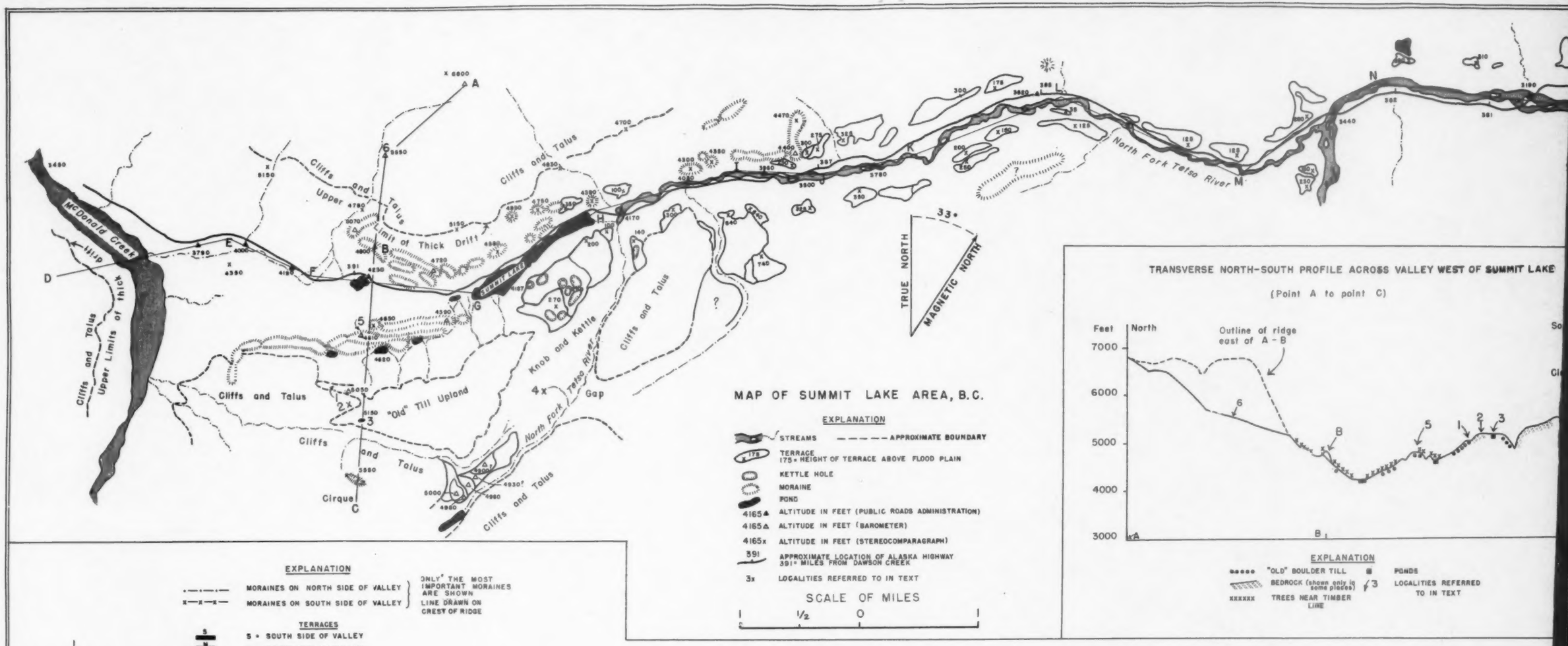
The terraces between Summit Lake and the Tetsa River (Pl. 4) are underlain by well-sorted sand and gravel, with the former predominating. They support a mixed forest of lodgepole pine and white spruce under which is a well-developed podzolic soil profile. The terraces along the Tetsa River east of the lake are underlain by crossbedded gravel and sand. The higher terraces are indented by kettle holes, and the lower ones show old stream channels.

Near Summit Lake the somewhat rounded pebbles in the terrace gravel are predominantly limestone or sandstone in varying proportions. Where limestone pebbles are abundant, those at the surface commonly show evidence of solution on their tops, and have a layer of secondary lime beneath them that is a quarter of an inch or more thick. In places the surface soil effervesces freely with dilute hydrochloric acid.

Muncho Lake.—Small cirques whose floors are 1500 to 2000 feet above lake level are visible on many of the higher peaks west of Muncho Lake and the upper portion of the Trout River Valley. Time was not available to visit any of them. No cirques were observed on the mountains east of the lake where peaks are probably not as high as those on the western side. Several large U-shaped valleys enter the Muncho Lake-Trout River valley from the west.

Small masses of cross-bedded sand and gravel are present along the east side of the lake between the great alluvial fans (Pl. 7, fig. 2; Fig. 6). Along the lower slopes just above the southwestern shore of the lake are some north-trending ridges that may be lateral morains built by an ice tongue in the Muncho Lake basin.

A high gravel terrace rises about 200 feet above the northern end of the lake. There is a lower terrace with kettle holes near the point at which the lake overflows into the Trout River. The high terrace extends down the Trout River Valley for 15 to 20 miles. About 6 miles north of the lake, the high terrace consists of bouldery gravel that contains mostly limestone boulders



MAP OF SUMMIT LAKE AREA, B.C.

EXPLANATION

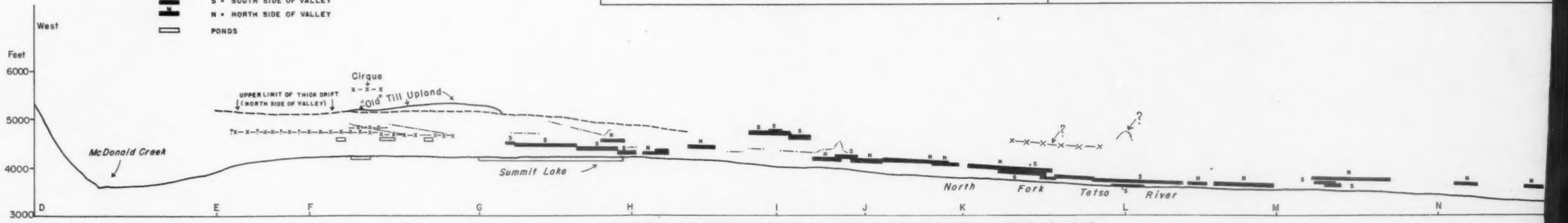
- STREAMS — — — — APPROXIMATE BOUNDARY
- TERRACE, 175 = HEIGHT OF TERRACE ABOVE FLOOD PLAIN
- KETTLE HOLE
- MORAINE
- POND
- 4165▲ ALTITUDE IN FEET (PUBLIC ROADS ADMINISTRATION)
- 4165△ ALTITUDE IN FEET (BAROMETER)
- 4165x ALTITUDE IN FEET (STEREOCOMPARAGRAPH)
- 391 APPROXIMATE LOCATION OF ALASKA HIGHWAY, 391 = MILES FROM DAWSON CREEK
- 3x LOCALITIES REFERRED TO IN TEXT

SCALE OF MILES

1 1/2 0 1

TRANSVERSE NORTH-SOUTH PROFILE ACROSS VALLEY WEST OF SUMMIT LAKE

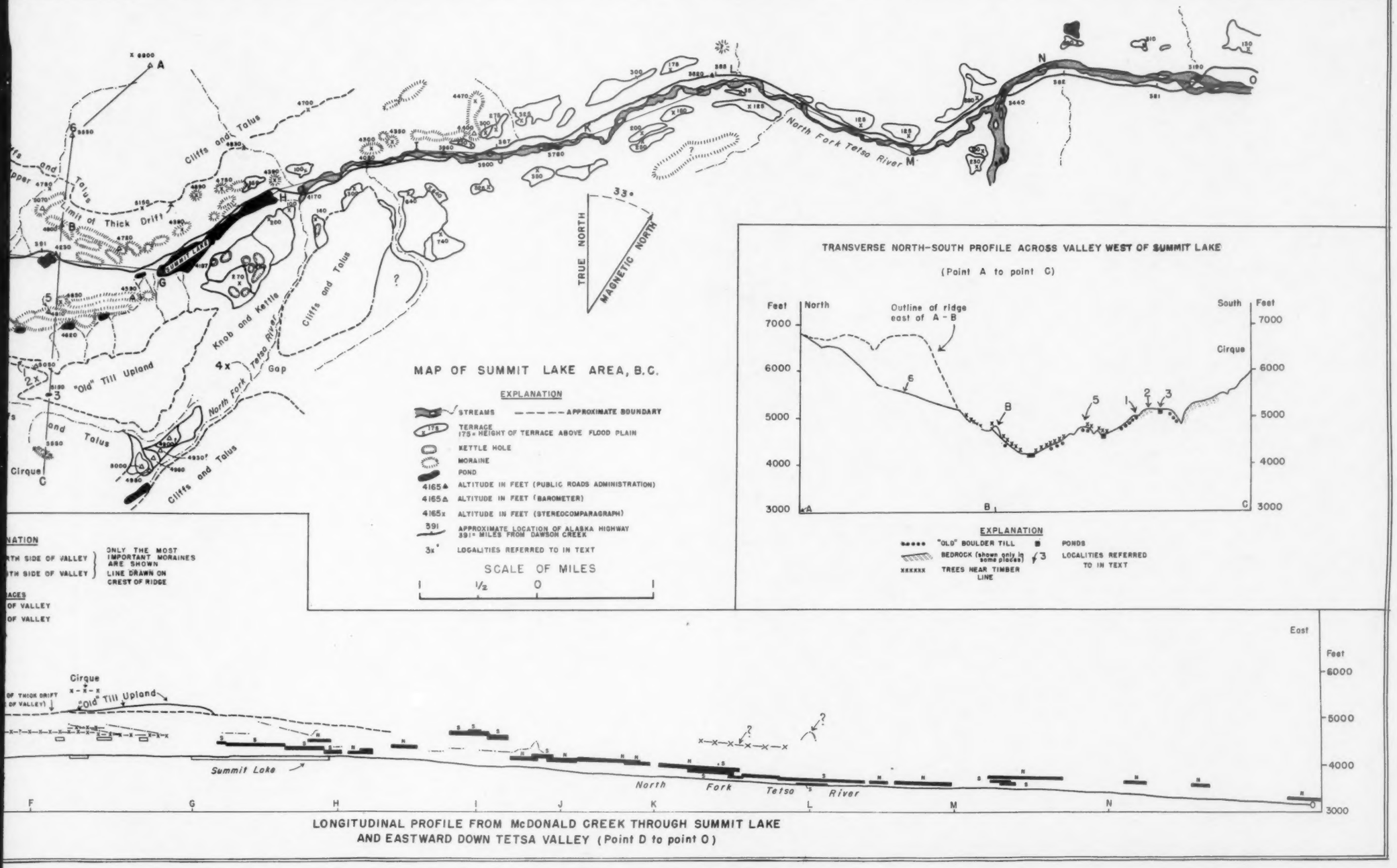
(Point A to point C)



LONGITUDINAL PROFILE FROM McDONALD CREEK THROUGH SUMMIT LAKE AND EASTWARD DOWN TETSA VALLEY (Point D to point O)

MAP AND PROFILES OF SUMMIT LAKE AREA, BRITISH COLUMBIA

Features shown on map are projected into line of profiles. Base map compiled from aerial photographs taken by U. S. Army Air Force, Alcan Highway Project, 1942; and Alaska Highway Plans, U. S. Public Roads Administration. Location and scale approximate.



LONGITUDINAL PROFILE FROM McDONALD CREEK THROUGH SUMMIT LAKE AND EASTWARD DOWN TETSA VALLEY (Point D to point O)

MAP AND PROFILES OF SUMMIT LAKE AREA, BRITISH COLUMBIA

Features shown on map are projected into line of profiles. Base map compiled from aerial photographs taken by U. S. Army Air Force, Alcan Highway Project, 1942; and Alaska Highway Plans, U. S. Public Roads Administration. Location and scale approximate.

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and rests on clean gray sand. Nearly 100 feet of this material is exposed. To the north the grain size of the material beneath the terrace decreases, and more and more fine sand is present.

river extends northward to the Liard, and is underlain by gravel and sand that shows very irregular stratification in some places (Pl. 7, fig. 3).

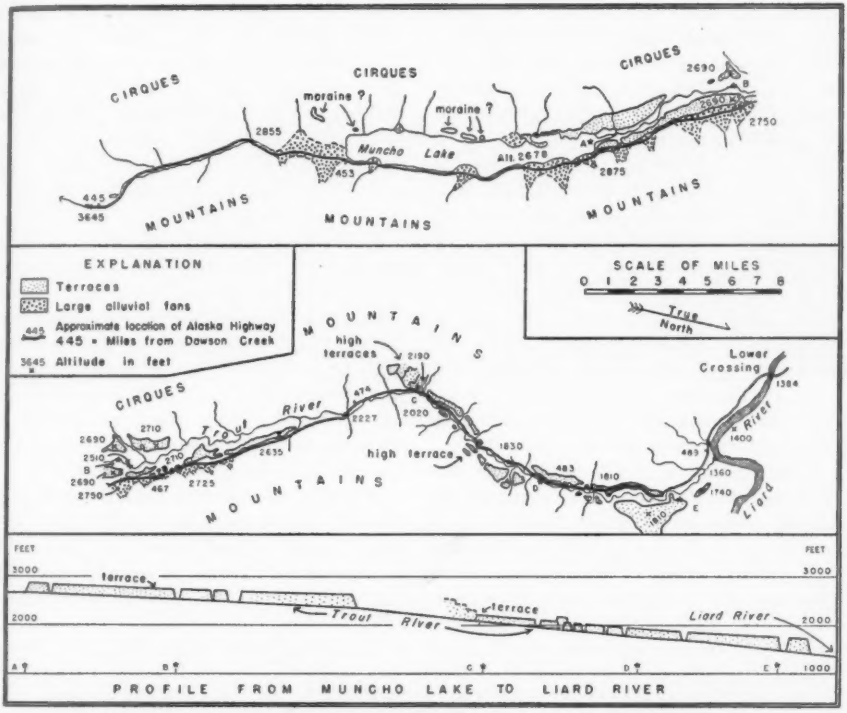

FIGURE 6. MAP AND PROFILE OF MUNCHO LAKE AREA, BRITISH COLUMBIA

Base map compiled from aerial photographs taken by U. S. Army Air Force, Alcan Highway Project, 1942; and Alaska Highway Plans, U. S. Public Roads Administration. Location and scale approximate.

About 14 miles north of the lake, the terrace is underlain by light-gray to white silt. This silt forms massive layers from 1 to 3 feet thick that are separated by thin seams of clay, and effervesces vigorously with dilute hydrochloric acid. A thin band of sub-angular pebbles as much as an inch in diameter was interbedded with the silt. This silt is overlain by a few feet of silty gravel that may be alluvial fan debris washed down from the highlands to the east. The silts are visible on both sides of the Trout River and underlie the high terrace that extends northward down valley to a point about 21 miles north of the lake. From approximately this point, a gravel terrace about 100 feet above

Liard Valley

Summary.—The upper Liard basin is characterized by extensive outwash plains and valley trains. All the glacial deposits are only slightly weathered and are probably late Wisconsin. How continuous the outwash plains originally were across the Liard Valley, or where, at any given time, the terminus of the Liard glacier was located is not known. Probably the outwash plains formed throughout a considerable interval of time, some before and some after the deposits near Watson Lake.

The outwash plains surrounding Watson Lake (Fig. 7) were built at some point during

deglaciation when the lake basin was still filled with ice. During this interval, drainage may have been eastward from the lake basin toward the Liard, rather than westward as at present. East of the lake are numerous gravel ridges associated with small outwash plains and kettle lakes, all of which suggest the break-up of ice into many detached blocks with sand and gravel deposited around them. At a slightly later time a tongue of active ice in the Liard Valley formed the gravelly moraine at the base of the basalt cliffs on the northeast side of the river north of the Upper Crossing. The ancient blockfield at the base of the basalt cliffs, between them and the moraine, was formed presumably when the ice stood along this moraine, and frost weathering of the low basalt cliffs was more active than it is at present. The upland forest partly covers the blockfield, indicating rather definitely that the latter is not being added to at present (Pl. 8, fig. 2). The formation of the blockfield between the lateral moraine and the basalt cliffs may represent a significant time interval.

In late glacial and post-glacial time, the Liard and its tributaries have dissected the outwash plains of the upper Liard basin to form a flight of river terraces distinguished from outwash terraces by the absence of kettle holes, the presence of abandoned stream channels on their surfaces, and a somewhat finer texture than characterizes many of the pitted outwash plains. Some of the terraces are nonpaired and rock-defined. The Liard is now deepening its channel slightly and widening its flood plain.

Terraces.—Remnants of a broad pitted outwash plain 200–300 feet above the Liard extend from Iron Creek to Watson Lake, about 60 miles. This outwash plain is underlain by well-stratified sand and gravel, and in some places is covered with low ridges of fine-grained sand; these may be old sand dunes now overgrown with lodgepole pine. Near Lower Post, about 15 miles southeast of Watson Lake, the outwash sand contains boulders of lignite (Pl. 8, fig. 1) that apparently floated in the streams that deposited the sand; these boulders apparently were derived from lignites of Tertiary age exposed along the Liard in the vicinity (Camsell and Malcolm, 1919; Lord, 1944).

At the Upper Crossing, near Watson Lake, the Liard is bordered by a sand and gravel terrace that rises about 100 feet above the river. This surface apparently is a post-glacial river terrace. Southwest of the Liard extensive outwash plains of gravel and sand rise about 400 feet above the river at a point several miles to the southwest. These plains are dotted by numerous kettle lakes and are surrounded by low hills of bouldery till and bedrock.

Watson Lake.—Watson Lake is bordered on a part of its northern and eastern sides by a sandy, pitted outwash plain (Fig. 7) about 50 feet above the level of the lake (altitude 2235 feet). There are also some gravel deposits above the level of the outwash plain.

The rolling upland around the lake is covered by a sandy till. The till contains an abundance of boulders of vesicular basalt together with quartzite, a medium grained, gray granite, diorite (?) porphyry, pegmatite (?), slate, and sandstone. The upland east of Watson Lake

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PLATE 5.—ALBERTA PLATEAU NEAR MUSKWA RIVER AND MORAINES NEAR SUMMIT LAKE, BRITISH COLUMBIA

FIGURE 1. REMNANTS OF ALBERTA PLATEAU NEAR FORT NELSON

Looking eastward across Muskwa Valley from point near Steamboat Mountain

FIGURE 2. MORAINES OF LAST GLACIATION IN SUMMIT PASS

View looking westward toward MacDonald Creek. Remnants of old upland surface in center background.

PLATE 6.—"OLD" TILL UPLAND NEAR SUMMIT LAKE, BRITISH COLUMBIA

FIGURE 1. VIEW LOOKING SOUTHWARD ACROSS SUMMIT PASS FROM MOUNTAIN SLOPE NORTH OF THE HIGHWAY

Smooth gullied slopes of "old" till upland in middle distance. Light streaks are gravel deposits formed during torrential rains of early July 1943. Cirques on peak in right background.

FIGURE 2. SMOOTH ROLLING SURFACE OF "OLD" TILL UPLAND SOUTHWEST OF SUMMIT LAKE

Pond (Pl. 4, Loc. 3) in foreground is floored with stone rings. Steep slope of post "old" till valley at right. Photograph taken July 1943.

FIGURE 1

FIGURE 2

ALBERTA PLATEAU NEAR MUSKWA RIVER AND MORAINES NEAR SUMMIT LAKE, BRITISH COLUMBIA

FIGURE 1

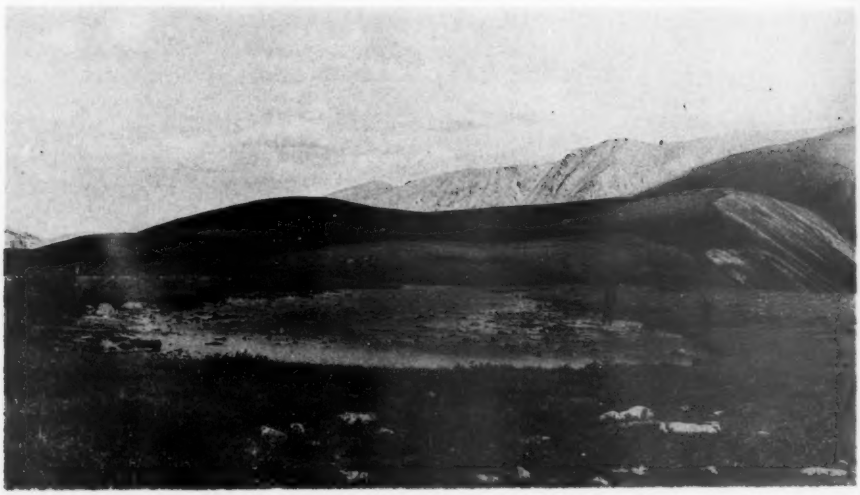

FIGURE 2

"OLD" TILL UPLAND NEAR SUMMIT LAKE, BRITISH COLUMBIA

FIGURE 1

FIGURE 2

FIGURE 3

DRIFT IN SUMMIT PASS AND IN MUNCHO LAKE-TROUT RIVER AREA, BRITISH COLUMBIA

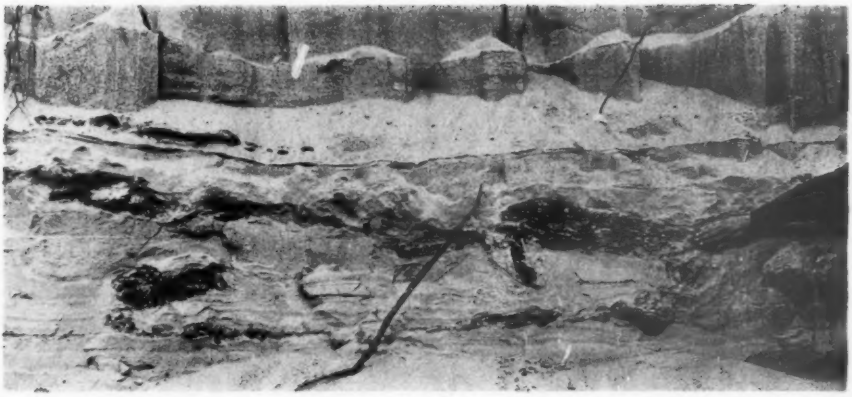

FIGURE 1

FIGURE 2

OUTWASH DEPOSITS IN LIARD VALLEY AND ANCIENT BLOCKFIELD NEAR WATSON LAKE

shows a complex system of abandoned stream channels, some of which drain down to lake level whereas others drain eastward toward the Liard River.

South and west of Watson Lake, the surface of the upland is apparently underlain by basaltic lava flows that form the edge of the plateau above the Liard. The basalt is overlain by pebbly till and sand.

Cassiar Mountains: Rancheria-Swift-Morley River Drainage

The principal stream valleys in the Cassiar Mountains are floored with drift, chiefly sand and gravel, that is dissected to form terraces along the valley sides. The deposits were laid down presumably in front of and around wasting valley glaciers that flowed east and west from the Continental Divide. These glaciers originated in the cirques near the divide and in the mountains north of Pine Lake. The glacial deposits have been dissected only slightly by the Rancheria and Swift rivers. No morainic deposits were seen. On the uplands between the Liard and Rancheria rivers, the till is a gray bouldery silt containing many basaltic fragments. Boulders of granitic rock are abundant in the drift elsewhere except near Teslin where the bedrock is schistose.

The drift is slightly weathered and is probably of late Wisconsin age. However, Lord (1944, p. 14) states that . . . "Near Mile 94E (along the Rancheria about 11 miles east of the Continental Divide) basalt was reported by an assistant to rest on drift." In the Tuya-Teslin area about 60 miles to the south, Watson and Mathews (1944) have described a series of volcanic rocks that were deposited in part before and in part after a time of glaciation. There is no evidence of volcanic activity since the latest glaciation.

Cirques are most numerous on the northern and eastern sides of the high peaks, and a few contain masses of ice on their headwalls.

Teslin Mountains: Teslin Lake to Whitehorse

Summary.—From Johnson's Crossing to Whitehorse, the Teslin Mountains are characterized by extensive terraces of gravel and sand along the valley sides. In general, the terraces are adequately explained as the result of deposition in ice-marginal streams and lakes, both of which need not have been very extensive at any one time. The presence of wasting ice is indicated by kettle holes, by the great debris-free lake basins, and by the presence of knob-and-kettle topography in valleys below surrounding terrace remnants (Bostock and Lees, 1938). It does not seem necessary to postulate any very large lakes, such as "Lake Yukon," for the origin of these deposits (Russell, 1890; Dawson, 1891; Spurr, 1898).

Most of the large lakes lie south of latitude 60°30′ (Pl. 2). Perhaps the lakes indicate a retreatal stage when the lake basins were filled with ice and terraces were formed, after which the ice wasted away without leaving any marked morainal deposits or terraces. The ap-

PLATE 7.—DRIFT IN SUMMIT PASS AND IN MUNCHO LAKE-TROUT RIVER AREA, BRITISH COLUMBIA

FIGURE 1. PILLAR OF "OLD" TILL ON SOUTH SIDE OF SUMMIT PASS
Boulders and pebbles cemented by calcium carbonate. MacDonald Creek in background.

FIGURE 2. GRAYISH-WHITE SAND AND GRAVEL CEMENTED BY CALCIUM CARBONATE
Exposure in road cut on east side of Muncho Lake

FIGURE 3. BOULDERY GRAVEL AND SAND EXPOSED IN ROAD CUT IN TROUT RIVER VALLEY ABOUT 5 MILES SOUTH OF LIARD RIVER

PLATE 8.—OUTWASH DEPOSITS IN LIARD VALLEY AND ANCIENT BLOCKFIELD NEAR WATSON LAKE

FIGURE 1. FLUVIATILE SAND UNDER HIGH OUTWASH TERRACE ALONG LIARD RIVER NEAR LOWER POST, BRITISH COLUMBIA
Dark masses are fragments of lignite

FIGURE 2. BLOCKFIELD ON TERRACE ABOVE LIARD RIVER NORTHEAST OF THE UPPER CROSSING
Lichen-covered boulders of basalt derived from low cliffs to right (outside of photograph). Blockfield in part overgrown by forest of spruce and aspen.

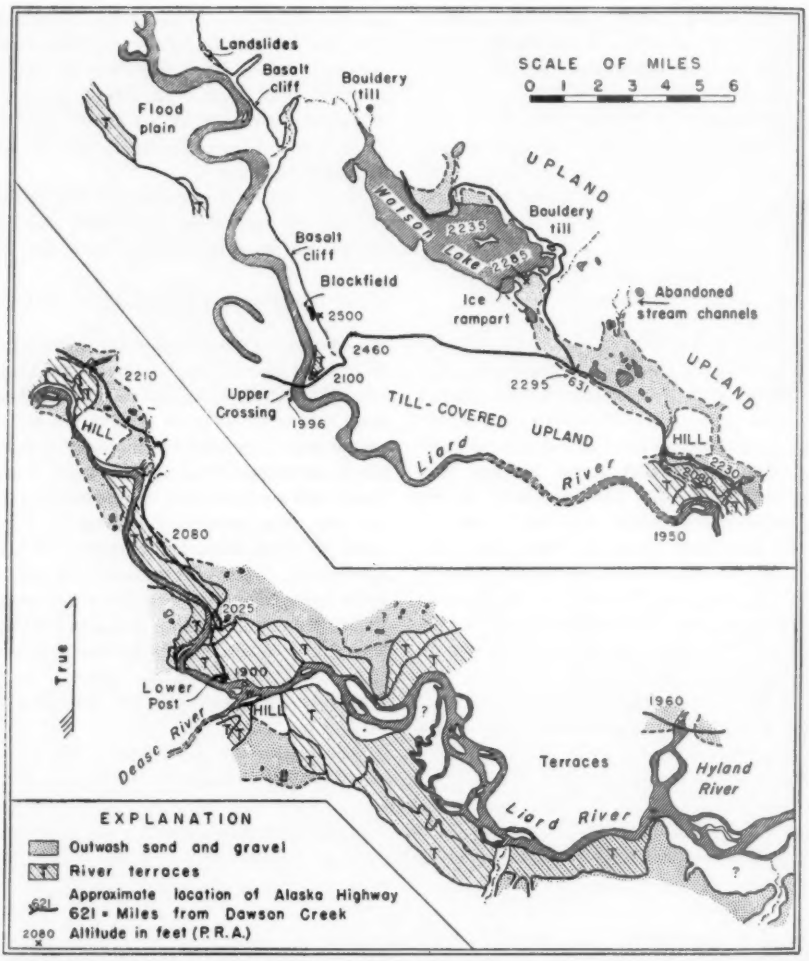

FIGURE 7. MAP OF AREA FROM WATSON LAKE, YUKON TERRITORY, TO LOWER POST, BRITISH COLUMBIA
Base compiled from aerial photographs taken by U. S. Army Air Force, Alcan Highway Project, 1942; and Alaska Highway Plans, U. S. Public Roads Administration. Location and scale approximate.

parently extensive silt deposits of the Teslin River Valley suggest a considerable period of deposition while the Teslin Lake basin was ice-filled. The outwash plain north of Little Atlin Lake and east of Marsh Lake and the terraces along the Lewes River suggest an interval of time when the basin of Marsh Lake was ice-filled.

Cirques.—Cirques are present on the northern and eastern sides of the higher peaks in the region from Teslin to Whitehorse. No cirques are visible on the mountains bordering Teslin Lake from Nisutlin Bay northward. However, cirques are visible on Dawson Peak on the west side of the lake just south of latitude 60°, and on the higher peaks in the Coast Ranges farther south. Cirques are also present on the peaks north of Squanga Lake and on the higher peaks south of the Highway from the north end of Teslin Lake westward to Little Atlin Lake, especially in the vicinity of Mount White. Small glaciers are present in cirques on the north side

of Mount Montana south of Carcross (Cockfield and Bell, 1926).

Terraces near Teslin Lake.—Varved sediments underlie a terrace that is about 80–100 feet above the northeast shore of Teslin Lake and extends from Brook's Brook to Johnson's Crossing, a distance of about 14 miles. About 4 miles southeast of Johnson's Crossing, 50 feet of lacustrine sediments are exposed. Alternating layers of gray silt and fine-grained sand, from 1–8 inches thick, are separated by thin seams of gray clay a fraction of an inch thick. The individual beds are thinner near the top where ripple marks are visible in the fine-grained sand. In the lower part of the section, the material is finer-grained. A few lenses of pebbly gravel are visible. Two terraces on the west shore of the lake near its outlet are apparently underlain by similar material. The lake beds extend far down the valley of the Teslin River to the northwest (Hayes, 1892; Lees, 1936). Teslin Lake is apparently dammed at least in part by glacial drift.

Little Teslin Lake, about 2 miles west of Johnson's Crossing, is completely surrounded by an extensive pitted outwash plain of sand and gravel that rises more than 300 feet above Teslin River. A rather flat sand plain dotted with kettle ponds extends westward from the lake for about 2 miles.

Beaches near Ten Mile Creek.—A small gravelly point juts out into Teslin Lake about 10 miles northwest of Teslin near Ten Mile Creek (Mile 809). Fine-grained pebbly gravel and coarse-grained sand are exposed in this point. Foreset beds dip to the south. The surface of this point is beveled by two narrow benches 7 and 17 feet above well-defined highwater mark that was 6 to 8 feet above lake level when visited on August 22, 1943. The benches, covered with slab-shaped pebbles, may possibly be old beaches and represent a higher stand of the lake. Artifacts were included in the sand and gravel of the higher bench (Johnson, 1946).

Squanga Lake and Valley North of Mount White.—About a mile southeast of Squanga Lake, a high gravel terrace about 150 feet above the lake is underlain by bouldery gravel and sand. To the south are several abandoned stream channels that apparently drain northward to the southern edge of the terrace. Abandoned channels are also abundant on the south side of the valley between Squanga and Little Teslin lakes.

The valley that extends from Squanga Lake westward to Little Atlin Lake is floored by intricate knob-and-kettle topography that contains numerous small lakes. Remnants of gravel terraces line the valley sides in some places. The western end of this valley, north of Mount White, is a U-shaped trough more or less free of glacial debris. The till exposed on the north side of the valley contains a high percentage of pebbles and boulders, angular or slightly faceted and soled. Granitic rocks, porphyry, marble, and sandstone are present. Some of the granitic pebbles are soft, are stained with limonite, and could be cut through with a shovel. The upper foot of the till is brownish and the calcareous sandy and silty matrix below is white but grades into gray at a depth of about 8 feet. Other sections show about 5 feet of loose pebbly silt. Silt in the upper 2 to 3 feet is brownish; in the lower 3 feet it is white, has a faint cleavage parallel to the ground surface, and effervesces vigorously with dilute hydrochloric acid. At a depth of about 5 feet there is slightly weathered, brownish, pebbly, silty till.

Marsh Lake.—The valley southeast of Marsh Lake and north of Little Atlin Lake is floored with sand and gravel that form a pitted outwash plain that slopes gently to the northwest, to a point about 300 feet above Marsh Lake. A chain of kettle ponds indents the outwash plain near its southern end and, to the north, a small stream has carved in these deposits a narrow canyon to depths of nearly 150 feet. The material is predominantly gravel, in places overlain by 2 or 3 feet of sand. The gravel is leached of calcium carbonate to depths of 2 or 3 feet, where there are white deposits of secondary calcium carbonate, characteristic of both the gravel and the till deposits of this region. Some of the granitic pebbles in the leached zone are soft and crumbly. The gravel is slightly stained by limonite.

Varved silt and clay, overlain by 2 to 3 feet of fine sand, are exposed at several points along the Highway. An exposure located about 12 miles southeast of McClintock River and about 200 feet above Marsh Lake showed 10 feet of varved sediment. The upper varves are less than an inch in thickness, but the lower are from 2 to 3 inches thick.

Lewes River to Whitehorse.—Terraces extend more or less continuously from Whitehorse southeastward to Marsh Lake, a distance of about 25 miles. At Whitehorse the terrace is underlain by about 200 feet of varved sediments that consist of light-gray to white, fine-grained sand and silt with thin layers of clay. Near the base of the section the sand and silt layers are from 1 to more than 3 feet thick, whereas toward the top the beds are somewhat thinner. The stratification is more or less regular, and no coarse sand and gravel were seen except a layer of fine-grained gravel near the top of the bluff. Large kettle lakes indent the terrace on the east side of the river.

Southward from Whitehorse the Highway is located near the western edge of a terrace some distance west of the river. Along the road are outcrops of bouldery gravel, sand, and silty till. The gravel underlies a terrace that is apparently slightly higher than the one formed by the varved sediments at Whitehorse. Study of aerial photographs suggests that the terrace underlain by varved sediments is more or less continuous for about 18 miles to the southeast, as far as the Lewes River bridge. South of the bridge, higher terraces rise several hundred feet above the river and apparently extend southeastward on the west side of the river to a point some distance south of the north end of Marsh Lake. Knolls and ridges of cross-bedded gravel just east of the bridge show forest beds dipping to the north.

Sand dunes near Carcross.—At the north end of Lake Bennett, an area of both modern and ancient sand dunes lies north and west of Carcross just east of Watson River. The dunes extend northward from the lake shore up a gentle south-facing slope for about a mile and a half. Irregular dune ridges and blow-outs extend north and south at right angles to the lake shore and are now partly overgrown with lodgepole pine. To the north, large areas of free sand rise up a south-facing slope to about 150 feet above the lake. On the north side of the dune field ridges of sand are separated by blow-outs in which there are tree stumps and small outcrops of bedrock. A more or less horizontal stratification is visible in the dune ridges on the sides of the blowouts.

The dunes are made of fine-grained gray sand. The sand ridges between the blow-outs are partly covered by a stunted growth of aspen, willow, and spruce, and are dotted by small grass-covered hummocks with streamlined ridges of sand on their lee (north) sides. The sand from the blow-outs is advancing northward up the slope. Where blow-outs occur, the dunes are building up slope; elsewhere, the relation of sand to vegetation is more or less stable.

The form of the dunes suggests that they were and are being built by southerly winds, as the sand is presumably derived from the sandy beach at the north end of Lake Bennett. This beach sand may have been carried by long-shore currents from the delta of Watson River just to the west. Dune formation probably began under periglacial conditions when Lake Bennett was still partly filled with ice, and has continued at a somewhat slower rate to the present day.

Sand dunes near Little Teslin Lake.—Just east of Little Teslin Lake are small knolls and ridges of fine-grained brown sand, probably old dunes now overgrown by lodgepole pine. The stratification is irregular. A few slightly wind-polished stones were found at the top of the underlying gravel.

Intensive Frost Action and its Effects

Introduction

Deposits, soil structures, or land forms, the product of intensive frost action and of related processes, characterize landscapes in far northern regions. Such products are numerous along the Alaska Highway (Table 3) and are intimately related to the occurrence of frozen ground and the distribution of vegetation. West of Whitehorse (Fig. 3) are extensive areas of permafrost (Muller, 1945) but, to the southeast, the ground is perennially frozen only in scattered localities (Johnston, 1930; Smith, 1939, Pl. 14). Soil structures and micro-relief features are more extensive and more varied west of Whitehorse than to the southeast. A second notable characteristic of surface features that result from present-day frost action is their absence or limited occurrence in forest-covered areas. Most of northern British Columbia and southeastern Yukon is forested and, as a result, frost action is now largely confined to areas

TABLE 3.—TABULAR SUMMARY OF FROST PHENOMENA ALONG SOUTHERN PART OF ALASKA HIGHWAY

Frost phenomena	Active or ancient	Localities where observed	Character of terrain where frost phenomena occurs		
			Nature of surface material	Water conditions	Type of vegetation
Permafrost	—	Widespread west of Whitehorse, and in scattered areas to the southeast	Absent in sand and gravel (at least near the surface); present in most other types of material	Poor to medium drainage	Alpine tundra, forest or bare ground
Congelifraction of bare rock slopes and summits	Both	Widespread; most active at high altitudes; most common in jointed rocks	Rock	Most active on damp surfaces	Sparse
Blockfields	Ancient	Probably widespread; occur on gentle to steep slopes, generally below timber line	Boulders	Well-drained	Sparse to covered with forest
Stone rings and stone stripes	Active	Widespread above timber line, on level land to steep slopes; occur in forested areas only under special conditions of vegetation and drainage	Till, congelifracts, and rarely on gravel	Most active in soils with abundant water	Scattered patches of alpine tundra or barren surfaces
	Ancient	Widespread on level ground to steep slopes; generally below timber line	Till or congelifracts	Usually well-drained	Sparse to covered with forest
Terraces and related forms (turf-banked terraces, solifluction terraces, terracettes, etc.)	Active	Widespread on moderate to steep slopes above timber line	Congeliturbate (solifluction debris), till	Most active in soils with abundant water	Usually some form of alpine tundra
	Ancient	Probably widespread on moderate to steep slopes, but usually obscured by forests	Congeliturbate (solifluction debris), till	Usually well-drained	Mostly under forest
Congeliturbation and other types of mass movement	Active	Widespread on moderate to steep slopes especially above timber line	All types of unconsolidated debris, especially those rich in fines	Poor to medium drainage	Alpine tundra and forest
	Ancient	Probably widespread on moderate to steep slopes below timber line	All types of unconsolidated debris, especially those rich in fines	Medium to well-drained	Sparse to covered with forest
Turf-covered hillocks	Active	Widespread on level to gently sloping land above timber line	Most common in unconsolidated debris rich in fines	Most active in soils with abundant water	Alpine tundra
Ice ramparts	Active	Lake shores on level to gently sloping land	Gravel and sand	At or near lake level	Forest

above timber line. Nevertheless, ancient and partly destroyed frost forms testify to former intensive frost action in areas now covered by trees.

From Whitehorse southeast to the Upper Crossing of the Liard River near Watson Lake, a distance of about 275 miles, the Highway was built over permafrost for only about 19 miles, or 6.5 per cent (Eager and Prior, 1945). Permafrost was noted also at a few isolated spots farther to the southeast (Pl. 3). Between Whitehorse and Big Delta (Fig. 8), 28 per cent of the road is built over ground that, at the time of construction, was classed as permafrost (Eager and Prior, 1945, p. 63).

Climatic conditions are favorable for frost

action. Annual and daily ranges of temperature are wide. The air is dry, and radiation cooling of the ground is pronounced at night. During spring and autumn, the temperature crosses the freezing point many times. Even during the hottest month of the year, July, freezing may occur anywhere at any time.

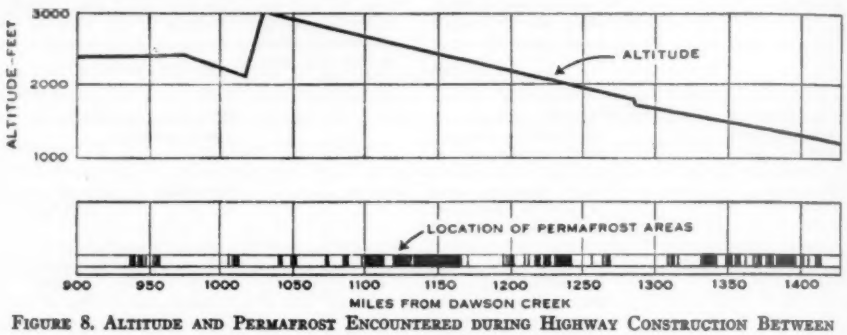

FIGURE 8. ALTITUDE AND PERMAFROST ENCOUNTERED DURING HIGHWAY CONSTRUCTION BETWEEN WHITEHORSE, YUKON, AND BIG DELTA, ALASKA
(Eager and Pryor, 1945)

Frost action cannot be vigorous unless a plentiful supply of water is available in the soil. In the region of the Alaska Highway, ground water is relatively abundant. The rate of evaporation is slight in spite of the dryness of the atmosphere; temperatures are low, and even that of the warmest month is moderate. The tundra plants are particularly effective in retarding evaporation from much of the land surface. The perennial freezing of much of the ground prevents underground drainage and is responsible for many swamps and a generally high water table.

Bryan (1946) has devised a terminology for the description of the processes and results of frost action. Some of the terms that he proposes are: *congelifraction* (Latin congelare: to freeze, fracture: to break) = frost-splitting or frost-riving; *congelifractate:* a body of congelifracts, or a mass of material of any grain size produced by congelifraction; *congeliturbation* (Latin congelare: to freeze, turbare: to stir up) = frost action including frost-heaving and differential and mass movements, and including solifluction, sludging, etc.; *congeliturbate:* a body of material disturbed by frost action = warp, trail, head, Coombe rock, Erdfliesse, Brödelerde, etc.; *cryoplanation* (Greek Kruos: icy cold; and plane through French from Latin planus) = land reduction by the processes of intensive frost action, i.e. congeliturbation, including solifluction and accompanying processes in translocation of congelifracts. This includes the work of rivers and streams in transporting materials delivered by the above processes.

A world-wide standard terminology for frost phenomena is desirable, but none is available at present. The term *permafrost* (Muller, 1945) has had remarkably wide acceptance. In this paper some of Bryan's terms have been very useful; namely *congelifract* and *congelifraction*, *congeliturbate* and *congeliturbation*, and *cryoplanation*. Edelman's terms "cryoturbate" and "cryoturbation" are also suitable, but have not been used in North America so far as the writer is aware (Bryan, 1948; Edelman, Florschütz, and Jeswiet, 1936; Hörner, 1950). Future discussions of terminology may produce other terms for these processes and their results.

Talus Slopes near Summit Lake

On either side of Summit Pass (Pl. 4), the till is thin or absent above altitudes of about 5000 feet. The upper slopes are covered with talus or composed of bare rock, most of which is flat-lying limestone strata with minor amounts of quartzite or sandstone. Boulders in the talus above timber line are predominantly limestone, many having surfaces etched by solution (Pl. 12, fig. 3). These solution-faceted boulders are in marked contrast to the angular fragments of limestone that form modern stone

rings (Pl. 10, fig. 1), described below, on the summits of the peaks north of the pass. The solution facets on the large boulders suggest that the latter are not moving down slope at present but have remained stationary long enough for the facets to develop. It is logical, therefore, to suppose that the talus is an ancient deposit, the result of previously active congeliturbation. However, such stability seems hardly possible. Small patches of alpine tundra occur even on steep talus slopes just below mountain summits. Some of the plants rest on relatively small, loose fragments of rock but have a remarkable root system. The roots extend upslope beneath a layer of fragments for more than 10 feet (Raup, 1947) and indicate at least 10 feet of downward movement of the fragments during the life of the plant. All that can be said with certainty is that the blocks of the stone rings on the summits appear to be more active at present than do large boulders in the talus on adjacent slopes.

The mountains north of the pass were not occupied by local glaciers in late Wisconsin time. The peaks and the intervening ridges have a transverse profile of two convex curves slightly rounded off at the top where they meet to form a broad ridge. The lower slopes are steep, and cliffs are visible in many places. Upward, the slope decreases slightly and loose blocks cover it all the way to the crest, which is a broad felsenmeer as much as 100 feet wide and slopes gently to either side. Intense congelifraction on ridge crests has apparently rounded them off slightly and mantled lower slopes with coarse rock fragments. Rounded divides are characteristic not only of peaks north of the lake but also of the divides in the foothills east of the Rocky Mountains.

Talus cones extend below the upper limit of the thick drift (Pl. 4). North of the Summit Pass, large masses of talus are covered by a scattered growth of white spruce and by an alpine scrub of dwarf birch that has not been disturbed by movement of boulders in recent years. This talus is composed of rounded and lichen-covered boulders of quartzite or sandstone. Some till-covered slopes between the cones show boulder stripes and breaks in the vegetation, the result of present-day congeliturbation and rain wash.

Blockfield near Watson Lake

On the northeast side of the Liard River west of Watson Lake, between a 100-foot river terrace and the basalt flows at the edge of the Watson Lake upland, is a gravelly moraine more or less parallel to the river (Fig. 7). This boulder-covered ridge is about 300 feet above the Liard River and is separated by a narrow swale about 30 feet deep from low basalt cliffs to the east. The swale is covered by an ancient blockfield of basalt boulders that extends east up a gentle slope to the base of the lava cliffs, which are only about 60 feet high. The blockfield is partially overgrown by a spruce and aspen forest and, in many places, is covered by a thick mat of mosses and lichens (Pl. 8, fig. 2). The relation of vegetation to the blockfield indicates that the latter is not being enlarged at present. It was probably formed when ice stood on the river side of the moraine, and frost weathering of the low basalt cliffs was more active than it is at present.

Stone Rings and Stone Stripes

Alberta Plateau.—Pebble rings of modern date were observed on a small deposit of bog iron (Hage, 1944, p. 20) on the side of a small gulch tributary to the Beatton River (Fig. 5, Loc. 2). The deposit is about 200 feet long and 75 feet wide. At two places in the area are cold bubbling springs, near one of which a miniature blockfield composed of red, slablike stones (Fig. 9) covers an area about 20 feet on a side. On this north-sloping surface are numerous stone rings and stone stripes about 6 inches across and 3 inches deep (Pl. 9, fig. 1). Between the stone stripes are small rock fragments from one-sixteenth to one-half inch in diameter, and a little sand. Below the stone rings there are 6 to 8 inches of red-brown sandy silt that rests on slabs of bog iron. When excavated on June 11, 1943, the ground was unfrozen to a depth of at least a foot.

At the lower (north) end of the stone stripes the blockfield is advancing into a scrub growth of dwarf birch and willow. At the upper (south) end, the root zone of the scrub is undercut to a depth of 6 to 10 inches. Some undercutting had taken place in the spring of 1943. No stone rings or stone stripes were found elsewhere in

the Beatton Valley. Apparently the soft bog iron and abundant water combine to make favorable conditions for the development of stone rings and to keep the vegetation in check.

at altitudes of nearly 7000 feet. The stone rings are from 2 to 4 feet in diameter and consist of angular fragments of limestone with centers composed of small chips and finer material that support small patches of alpine

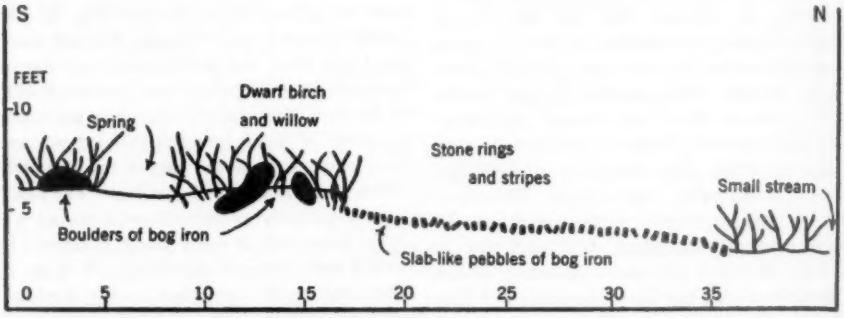

FIGURE 9. STONE RINGS AND STONE STRIPS ON SMALL DEPOSIT OF BOG IRON ALONG BEATTON RIVER, BRITISH COLUMBIA

A blockfield of angular and somewhat rounded, lichen-covered boulders is crossed by the army pioneer road on top of the plateau about 4 miles north of the Beatton River bridge and at an altitude of about 4000 feet (Fig. 5, Loc. 1). Sandstone boulders as much as 5 feet in diameter form rings whose centers consist of pebbles and a little sandy silt. A few small spruce trees and moss grow on the centers. Many slabs stand on end. The blockfield disappears under the surrounding black spruce forest (Pl. 9, fig. 2). The rings do not appear to be forming today, as indicated by their partial coverage by the forest, and by the roundness and lichen cover of the individual stones.

Summit Lake.—Stone rings are well developed on top of peaks north of Summit Lake tundra (Pl. 10, fig. 1). The fragments in the rings are markedly angular in shape, as compared with boulders in rings at lower altitudes, and show little evidence of solution. Bedrock is exposed in some places. The slopes immediately below the summits are covered with boulder stripes separated by masses of smaller, angular rock fragments. Tundra grows only in small patches between the fragments. The ridge crests between one summit and the next are rounded surfaces tens of feet in width. Boulder stripes extend over the crests and on either side of these ridges.

Stone rings are found on the "old" till upland south of the pass (Pl. 4, Loc. 1), at an altitude of just over 5000 feet. Here on an approximately level surface of lime-cemented boulder till are oval-shaped rings from 6 to 20

PLATE 9.—STONE RINGS IN ALBERTA PLATEAU, BRITISH COLUMBIA

FIGURE 1. SMALL STONE RINGS ON SURFACE OF DEPOSIT OF BOG IRON IN BEATTON RIVER VALLEY (*See* Figure 9)

FIGURE 2. ANCIENT STONE RINGS ON ALBERTA PLATEAU NORTH OF BEATTON RIVER

Angular and slightly rounded lichen-covered boulders of sandstone form a boulder pavement that is partly over-grown by surrounding spruce forest.

PLATE 10.—STONE RINGS NEAR SUMMIT LAKE, BRITISH COLUMBIA

FIGURE 1. STONE RINGS ON SUMMIT OF MOUNTAIN JUST NORTH OF SUMMIT LAKE

Altitude about 7000 feet. Rock is limestone. Centers of rings support alpine tundra

FIGURE 2. ANCIENT STONE RINGS ON TUNDRA-COVERED SURFACE OF "OLD" TILL UPLAND SOUTHWEST OF SUMMIT LAKE

(*See* Pl. 4, Loc. 1.) Rings are composed of slightly rounded lichen-covered boulders, chiefly of sandstone.

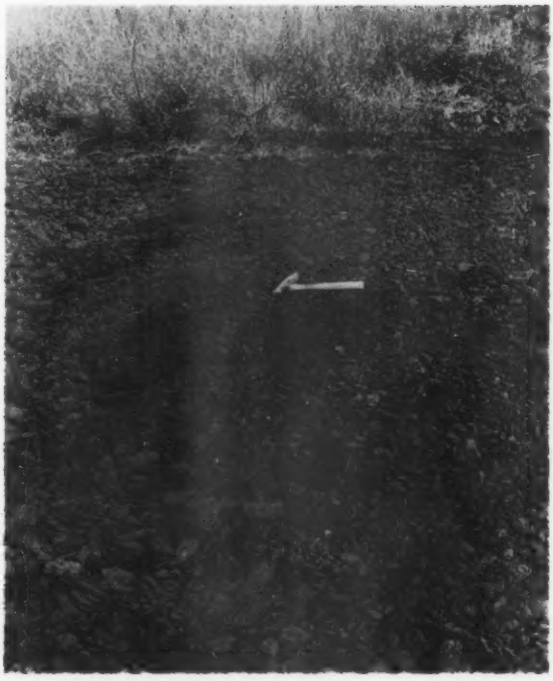

FIGURE 1

FIGURE 2

STONE RINGS IN ALBERTA PLATEAU, BRITISH COLUMBIA

FIGURE 1

FIGURE 2

STONE RINGS NEAR SUMMIT LAKE, BRITISH COLUMBIA

FIGURE 1

FIGURE 2

STONE RINGS NEAR SUMMIT LAKE, BRITISH COLUMBIA

FIGURE 1

FIGURE 2

FIGURE 3

STONE RINGS, STONE STRIPES, AND TALUS NEAR SUMMIT LAKE, BRITISH COLUMBIA

feet across. The rounded and lichen-covered boulders that form these rings are ½ to 2 feet in diameter and consist of limestone and sandstone. The boulder bands are from 2 to 6 feet wide (Pl. 10, fig. 2). The centers of the rings are covered by a growth of lichens, grasses, sedges, dwarf birch, and willow, and when first visited on July 13, 1943, were very moist.

Just south of these stone rings, limestone ledges about 100 feet high (Pl. 4, Loc. 2) are, breaking up into angular blocks of all sizes, from 6 feet in diameter to small chips. This talus forms irregular tongues and stripes that descend a north-facing 20° slope and invade the tundra-covered surface of the upland to the north. The limestone talus blocks, large and small, all show surfaces etched by solution and are largely free of lichens, although there are a few lichen-covered boulders among them. In two respects, lichen cover and degree of roundness, these talus blocks differ markedly from the boulders in the rings to the north. The limestone chips, slab like in shape, stand in the stripes with their long axes more or less vertical. The above-ground portion of the chip is etched by solution whereas the lower end in the ground is not. The stripes that descend the north-facing slope give every indication of having been in motion during the previous spring, although all movement apparently had ceased when they were first visited on July 13. No permafrost was present above a depth of at least 18 inches.

At first glance the rings on the upland (Loc. 1) appear to be older and more stable than the stripes to the south, as indicated by the roundness of the boulders in the former and their lichen cover. However, these are glacial boulders derived from the "old" till and were therefore partly rounded before they were subjected to congeliturbation. They are the more resistant, joint-free fragments of bedrock that were segregated by the glacier, whereas the angular limestone chips on the slope to the south are derived directly from bedrock and have not been subjected to any selective action. The lichen-covered surfaces of the glacially rounded boulders are probably more resistant to continued break-up than are the jointed talus fragments in the stripes. Nevertheless, the relation of these two soil structures to the mat of tundra vegetation, which is more disturbed on the talus slope than on the flat below, suggests that during the spring of 1943 there was more movement on this talus slope than in the rings on the upland.

On gentle slopes (less than 20°) in this vicinity large limestone boulders are breaking up in place into small chips. Some boulders have a small train of angular chips downslope from them for a distance of as much as 15 feet. Elsewhere the boulders have been almost completely disintegrated by congelifraction, and the stripes of limestone chips have at their heads only small fragments of the original boulders.

Near the headwaters of the Tetsa River, stone rings (Pl. 12, fig. 1) mantle the nearly level floor of a small valley above timber line at an altitude of about 4600 feet (Pl. 4, Loc. 4). The rings are made of boulders as much as 2 feet in diameter that surround centers of pebbly silty sand overgrown with alpine tundra. A

PLATE 11.—STONE RINGS NEAR SUMMIT LAKE, BRITISH COLUMBIA

FIGURE 1. STONE RINGS IN BED OF SMALL POND ON "OLD" TILL UPLAND SOUTHWEST OF SUMMIT LAKE

(*See* Pl. 4. Loc. 3.) For section through rings see Figure 10. Photograph taken September 1944. (*See* also Pl. 6, fig. 2.)

FIGURE 2. PART OF STONE RING SHOWING MINIATURE RINGS IN CENTER OF LARGER RING

For section through these rings see figure 10. Same locality as figure 1

PLATE 12.—STONE RINGS, STONE STRIPES, AND TALUS NEAR SUMMIT LAKE, BRITISH COLUMBIA

FIGURE 1. STONE RINGS NEAR HEADWATERS OF TETSA RIVER, SOUTH OF SUMMIT LAKE

(*See* Pl. 4, Loc. 4.) Rings are probably ancient forms exposed by stream erosion

FIGURE 2. ANCIENT STONE STRIPE AT TIMBER LINE SOUTH OF SUMMIT LAKE

Stripe is bordered by an alpine scub of dwarf birch and willow

FIGURE 3. SOLUTION-FACETTED LIMESTONE TALUS BOULDERS ON SLOPE SOUTHWEST OF SUMMIT LAKE

Location is just south of Loc. 1, Plate 4.

stream flows through the stone borders that pass under the tundra on either side of the channel. These rings may be forming now or they may be old, more or less stable forms once covered by tundra and now exposed by stream erosion.

turn is bordered by alpine tundra where grasses are more abundant than sedges (Fig. 10).

The ring of sedges suggests that the pond was once slightly more extensive than at present, or at least that the water-level has been higher

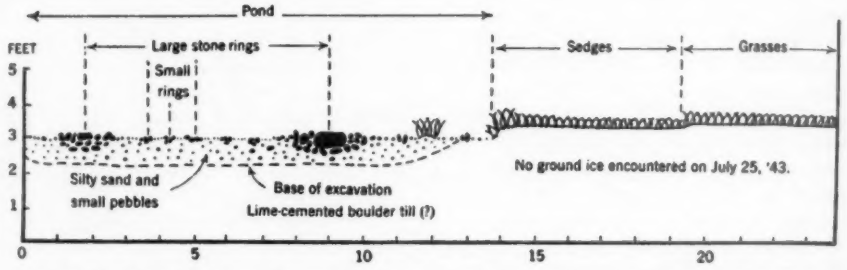

FIGURE 10. STONE RINGS ON BOTTOM OF SMALL POND ON "OLD" TILL UPLAND SOUTHWEST OF SUMMIT LAKE

Altitude about 5150 feet (Pl. 4, Loc. 3)

Some centers are turf-covered hummocks as much as 2 feet high and several feet across.

A small pond on the "old" till upland southwest of Summit Lake (Pl. 4, Loc. 3) is at an altitude of about 5150 feet. Stone rings (Pl. 11, fig. 1, 2) carpet the bottom of the pond which, on July 13, 1943, had a maximum depth of about 3 feet. On July 25 about half the bottom was exposed, and when seen by Sticht and Raup in September 1944, the whole pond was dry. The stone borders, from 1 to 2 feet wide, consist of angular and slightly rounded pebbles and cobbles, the majority from 4 to 8 inches in diameter. These slightly rounded stones are probably derived from the underlying "old" boulder till. The centers, 2–5 feet across, are carpeted with silty sand and small pebbles. The material in the centers of large rings is sorted into imperfect miniature stone rings that range from 3–6 inches in diameter (Fig. 10; Pl. 11, fig. 2). Near the edge of the pond, sedges grow on parts of the centers of the larger rings. A trench about a foot deep, excavated on July 25, 1943, across the exposed bottom of the pond, showed that the large stone borders extend to depths of 6 inches to 1 foot and are underlain and separated by silty sand that contains a few pebbles. No ground ice was encountered. The miniature stone borders in the centers of large rings are only 1 or 2 inches deep. The pond is surrounded by a sedge-covered flat, which in than it was on July 13, 1943. However, the sedges might indicate an advance of the tundra over surface of the pond, suggesting a decrease in intensity of congeliturbation that produced the large stone rings. The latter might therefore be in part relict forms. The miniature stone rings within the larger centers can be used to support this hypothesis. If the larger centers were actively heaved during 1943, and if stones were added to the larger rings, these large "active" centers should not be broken up into miniature rings. Perhaps the present frost cycle is sufficient to produce only the miniature rings. On the other hand, observations elsewhere indicate that miniature rings can form in one season (Denny, 1940). Therefore, the miniature rings may indicate merely that congeliturbation during 1943 did not add to the larger stone rings, which may have been formed in part during any previous season when frost heaving at this spot was more intensive than during 1943. For somewhat analogous relationships, Troll (1944, p. 619 ff.) suggests that large rings are the result of annual temperature changes that penetrate to considerable depth, and that miniature rings are due to daily temperature changes that affect only the upper inch or two of material at the surface.

Stone rings with tundra-covered centers occur in many places below timber line. Rounded, lichen-covered boulders form rings at the base

of some of the moraines on either side of the pass and in kettles on outwash terraces southeast of Summit Lake. The centers of the stone rings are covered with alpine tundra or with an alpine scrub of dwarf birch and willow.

Cassiar and Teslin Mountains.—Stone stripes of rounded, lichen-covered boulders are found on the slopes of many terraces along the Rancheria River and the bordering upland. In the Teslin Mountains, the peak north of Mount White (Pl. 2) is covered by a felsenmeer more or less overgrown by alpine tundra. Stone stripes made of somewhat rounded, lichen-covered boulders are separated by turf rich in lichens. There are also irregular masses of boulders surrounded by alpine tundra.

The boulder-covered slopes of this and neighboring peaks appear to be more stable than similar slopes above timber line near Summit Lake. It is difficult to make a direct comparison, nevertheless the stone stripes in the area near Mount White are apparently not forming today as actively as in the past. No pebble-covered pavements or breaks in the alpine tundra, structures so abundant near Summit Lake and indicative of active present-day congeliturbation, were seen on the peaks near Mount White.

Age of stone rings.—The growth form and distribution of alpine tundra around stone rings (or other related soil structures) show whether or not there has been movement of stones during recent years. Above timber line, some boulders have a tundra-covered ridge on their downslope side and a break or gap in the vegetation above them, indicating that the boulders have moved downslope since the last growing season. Also patches of bare mineral soil (pebble-covered pavements or turf-banked terraces) are associated with stone rings and stone stripes that have alpine tundra growing in the centers of the rings or between the stripes. The mineral soil in these bare patches is loose-textured and almost certainly was frost-heaved in 1943. However, in many places it was not possible to demonstrate that the boulders of the rings actually moved during 1943. The development of the alpine tundra is obviously related to the stability of the soil, but whether one type of alpine tundra is more indicative of stability than anothes is a problem for the botanist (Raup, 1947; Polunin, 1934–1935).

The problem of a possible differentiation between stone rings made of rounded, lichen-covered boulders and of angular, lichen-free boulders is a fundamental one. One possible explanation is that there are two ages of stone rings; that the rounded, lichen-covered boulder rings are old, their roundness being the result of weathering after the rings were formed, and the lichen cover developed after the boulders were actively moved by frost heaving. Stone rings made of angular, lichen-free boulders occur where present-day frost action is probably most active and vegetation scanty, as on mountain summits. However, other factors are involved. The rounded boulders are derived in many places from bouldery till or gravel, so the roundness may be an inherited characteristic.

Lichen growth is a slow process; the presence of a lichen cover over a boulder may indicate that the boulder has been in a stable position for a "long" time. However, it is obvious that many lichen-covered boulders are now moving rapidly downslope. Once a cover of lichen has grown on a boulder it is very difficult to remove. L. C. Raup states (personal communication) that crustose lichen becomes gelatinous, especially when wet, and that a boulder covered with such lichen may easily slide without scraping off much, if any, of the lichen crust. Therefore, the presence or absence of a lichen cover is not in itself a valid criterion for the stability of a boulder. However, boulders with a lichen crust on their upper or exposed surface, but absent below, suggest that the boulders have not been overturned during or since the time of lichen formation.

The age of stone rings in the alpine scrub zone or below timber line is difficult to determine. In general, the vegetation in and about such rings showed no evidence of disturbance by congeliturbation either during 1943 or during several years previous. The rings in kettle holes southeast of Summit Lake are on well-drained, sandy material where the water table and permafrost, if present, are at considerable depths, probably now too far below the rings to facilitate their disturbance by frost heaving. It may be supposed that they were formed when the permafrost table was close to the surface. Below timber line, temperature changes through the freezing point are not as frequent as above,

where the ground is not in shade or blanketed so long or so continuously with snow. The boulder rings on the Alberta Plateau between the Beatton and Sikanni Chief rivers are surrounded and partly overgrown by a spruce forest and do not appear to be forming at present. In contrast, the modern rings near the Beatton River apparently owe their occurrence to special edaphic conditions.

SUMMARY: The stone rings and stone stripes along the Highway present a more or less continuous series, from relict forms to rings that certainly are being actively formed at present. Some boulder rings in the Summit Lake area, especially those below timber line, and the boulder rings on the Alberta Plateau near the Beatton River are the result of processes that are no longer active, and possibly date from a period of intense frost action before the postglacial optimum. Each individual stone ring or group must be considered in relation to its various environmental factors, such as depth to ground water and to permafrost. No altitude can be given as the dividing line between ancient and modern stone rings.

In the Summit Lake area, there are two dominant types of soil structures: (1) pebble-covered pavements and turf-banked terraces (described in latter paragraphs); and (2) stone rings. In general, the vegetation around the first type showed obvious evidence of congeliturbation during 1943, whereas that around the second type did not, except for the stone rings on mountain summits. This fact suggests that, during the previous few years, frost action had been intense enough to produce pebble-covered pavements and turf-banked terraces, but not sufficient to enlarge previously existing stone rings.

Permafrost is much more widespread and occurs much nearer the surface in southwestern Yukon than in the region southeast of Whitehorse. In southwestern Yukon it aids the development of soil structures above timber line. The scattered areas of permafrost southeast of Whitehorse occur at depths of 5 to 10 feet and do not appear to be actively influencing the development of soil structures at present, except in a few localities. There is at least a suggestion that the areas of permafrost southeast of Whitehorse are relicts of a once more continuous body of perennially frozen ground that led to the development of soil structures in the past. Most modern soil structures in the region from Dawson Creek to Whitehorse do not appear to be definitely related to the presence of permafrost.

The data from the area along the Highway seem to indicate rather definitely that permafrost and related soil structures were much more widespread phenomena in the past than at present.

Terraces and Related Forms in the Summit Lake Area

At many localities above timber line in the Rocky Mountains, the tundra is broken by oval or elongated patches of bare mineral soil. These surfaces are covered with pebbles and silty sand in varying proportions, and range from a few inches to many feet in length. On some slopes of moderate declivity, they are elongated into stripes. On slightly steeper slopes, the bare patches or pavements form the tops of turf-banked terraces, which in some places form a series of steps. The upper surface of a pavement may be essentially flat or have a profile that is faintly convex upward. Some pavements are covered by miniature stone rings a few inches across. All these forms are underlain by an almost pebble free, silty sand that rests in turn on pebbly or bouldery till.

Pebble-covered pavements are found on the nearly level top of the moraines southwest of Summit Lake and on the "old till" upland (Pl. 4, Locs. 3 and 5). The alpine tundra of lichens, sedges, grasses, and other plants is interrupted by bare, pebble-covered pavements (Fig. 12) underlain by pockets of silty sand. The underlying material is a pebbly sandy till. The alpine tundra between the two pavements is underlain by a dark horizon rich in organic matter and 2 or 3 inches thick.

Elsewhere, the alpine tundra between bare pavements is underlain by pebbles and cobbles without any fine material. If the tundra is stripped off, pebble-covered pavements separated by bands of pebbles and cobbles are exposed, the whole resembling typical stone rings. This arrangement suggests that the alpine tundra has gained a foothold on the stone borders, whereas the pebble-covered pavements are frost-heaved sufficiently each year to pre-

vent tundra growth. The silty sand is perhaps more susceptible to daily frost heaving than are the stone borders (Troll, 1944). On the other hand, many boulder rings as, for example, at Locality 1, Plate 4, have tundra-covered centers surrounded by bare boulder rings.

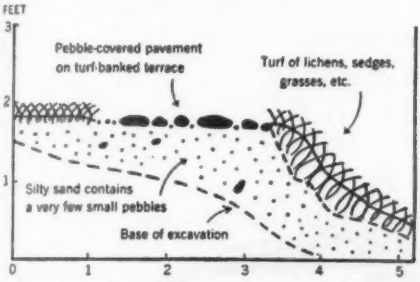

FIGURE 11. TURF-BANKED TERRACE NEAR SUMMIT LAKE, BRITISH COLUMBIA

Field sketch of relations exposed in trench. Locality is on "old" till upland southwest of Summit Lake.

Turf-banked terraces are abundant on slopes southwest of Summit Lake. Figure 11 shows a pebble-covered pavement on a turf-banked terrace above timber line at an altitude of about 4800 feet (Pl. 4, Loc. 5). The pavement is underlain by silty sand that contains a very few small pebbles. The tundra bank on the downslope side, at the right, is free of pebbles. Most turf-banked terraces are discontinuous along the slope. Most are not more than 15 feet long, many much less, and form a series of irregular breaks on tundra-covered slopes.

The pebbles on some pavements stand on tiny pedestals of silty sand a fraction of an inch in height, probably the result of erosion by impact of raindrops and by rain wash during the torrential rain of early July 1943. They emphasize that rain wash is a factor in the preservation and perhaps the development of pavements. The mat of alpine tundra on slopes above timber line is loose in many places, and there is probably considerably runoff through and under it.

Northwest of Summit Lake, near the headwater of a stream that drains westward to MacDonald Creek, several larger, turf-banked terraces occur on a south-facing slope at altitudes of about 5550 feet (Pl. 4, Loc. 6). The turf-banked terraces are up to 10 feet wide, 25 feet long, and as much as 5 feet high on their downslope sides. Terrace tops are covered with alpine tundra interrupted by several pebble-covered pavements. The turf-banked terraces are located on an 18° slope probably underlain by pebbly till. One somewhat imperfect terrace

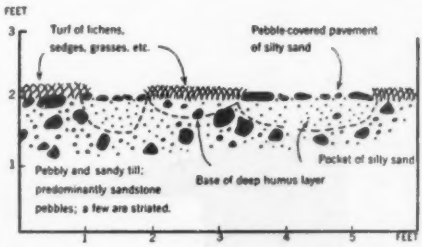

FIGURE 12. PEBBLE-COVERED PAVEMENTS NEAR SUMMIT LAKE, BRITISH COLUMBIA

Field sketch of relations exposed in trench. Locality is on "old" till upland southwest of Summit Lake.

extends horizontally about 100 feet. These larger turf-banked terraces resemble detritus benches in central Alaska (Taber, 1943). Many small, pebble-covered pavements and isolated boulders scattered over this tundra-covered slope showed evidence of frost heaving and downslope movement during 1943. However, the tundra on the larger turf-banked terraces did not show evidence of recent movement. This south-facing slope is the only locality where large turf-banked terraces were observed in the Summit Lake area.

Mass Movements in the Alberta Plateau

Slumps or debris slides (C. F. S. Sharpe, 1938, p. 65) are widespread along the Highway between Dawson Creek and Summit Lake, British Columbia. The walls of the Beatton Valley show evidence of mass movements, both topographic and in the underlying materials. The south-facing slope on the north side of the valley east of the army pioneer road is broken by elongated, aspen-covered knolls (Fig. 5, Loc. 9). The remainder of the slope is covered with mixtures of white spruce and lodgepole pine (Raup and Denny, 1950, Pl. 15, 16). These knolls are scattered at various altitudes along the slope and are composed of a brownish-gray, unsorted, sandy silt that contains angular

fragments of sandstone. On steep south-facing slopes along the army pioneer road north of the river (Fig. 5, Loc. 3), a compact clay till is overlain by loose, unsorted, sandy and silty material is exposed on the south side of the Beatton Valley and elsewhere. Such movement is not very active today on these wooded slopes, suggesting that mass movements by

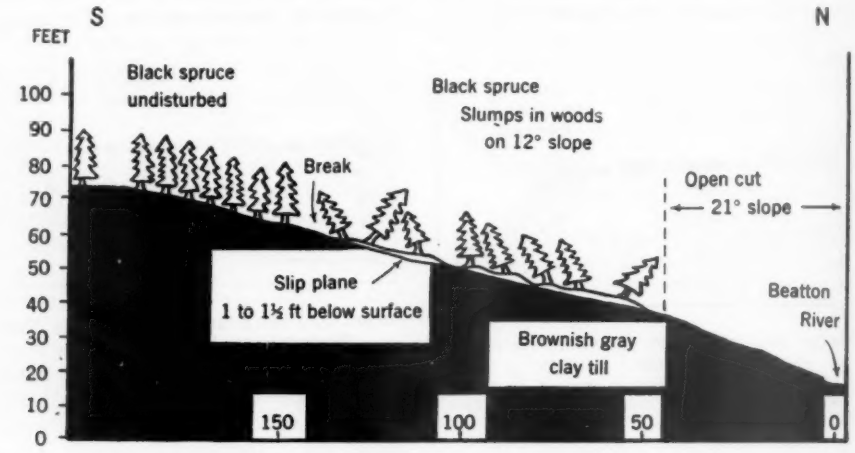

FIGURE 13. SLUMPS ALONG BEATTON RIVER, BRITISH COLUMBIA

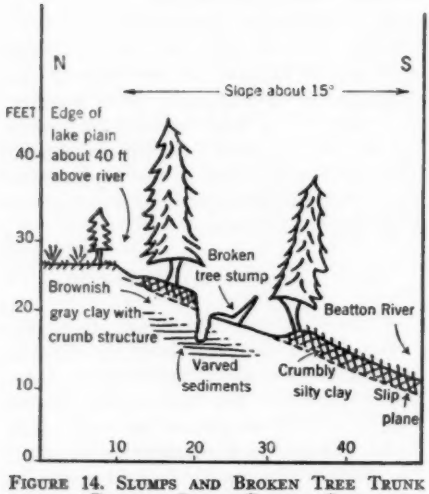

FIGURE 14. SLUMPS AND BROKEN TREE TRUNK ALONG BEATTON RIVER, BRITISH COLUMBIA (Fig. 5, Loc. 6).

material 4–6 feet thick that contains angular fragments of sandstone up to 6 feet in diameter. Apparently this material has been formed by downslope movement of till and angular rock fragments derived from near-by ledges. Similar congeliturbation were more active in the past then at present (Denny, 1938).

The bluffs along the Beatton River are covered in many places by slumps or debris slides that move down a slope that ranges roughly from 2° to 20° (Fig. 13; Fig. 5, Loc. 4). The slumps are 1 to 2 feet thick and include one or more trees whose roots aid in holding the clayey material of the slump together. Some trunks were split presumably by frost and dragged apart by the slumping movement (Fig. 14). Most slumped trees point upslope, but all possible inclinations were observed. The slumps range from a few feet to 100 feet long, but their width is much smaller. Uphill behind the slumps there is a curved or bow-shaped depression.

Some slumps are the result of undercutting by the river. However, slumps occurred during the spring of 1943 as indicated by the lack of vegetation in the uphill depressions on slopes not actively undercut. Most slopes on which there had been slumping during the spring of 1943 were steeper than those where the scars behind the slumps, when observed in June, were covered by vegetation that grew during the previous summer (notably *Petasites* sp., and *Equisetum arvense*). There is a complete grada-

TABLE

		ATHABASKA—GREAT SLAVE LAKE REGION	SOUTHERN MACKENZIE MOUNTAINS	ALBERTA	NORTHEASTERN BRITISH COLUMBIA		
		Development of forests on Cretaceous Plateaus (Raup, 1946)	Brintnell Lake, South Nahanni River (Raup, 1947)	(Bretz, 1943)	Alberta Plateau; Dawson Creek to Fort Nelson (Hage, 1944; Denny)	Summit Lake (Denny)	
RECENT		Black spruce—lodgepole pine ? Park-like white spruce			Small stone rings along Beatton River River terraces	Stone rings on mountain summits and small turf-banked terraces and related forms just above timber line River terraces	
	?						
	? POSTGLACIAL OPTIMUM	? Prairie	? Forest ? Tundra		? Prairie in Beatton Valley. Climate slightly milder than at present. No permafrost. ? Artifacts near Mill Creek	? Solution of boulders in talus above timber line	
?	?	Tundra					
PLEISTOCENE	WISCONSIN		Moraines of last glaciation Drift older than last glaciation	Altamont moraine (Late Wisconsin maximum) Abundant lakes Older moraines (Viking and older) formed by a pre-late Wisconsin ice advance	? Possible Altamont stage east of highway. Beatton Valley lake (?) Widespread permafrost and intense frost action. Boulder rings, solifluction on slopes of valleys Advance of Laurentide ice sheet to eastern edge of Rocky Mountains Lakes Advance of glaciers from Rocky Mountains	Moraines in cirques (5200–5800 feet), possibly post Wisconsin Widespread permafrost and intense frost action. Smoothing of surface of "old" till upland by solifluction North-south moraine 2 miles east of Summit Lake (4350 to 4400 feet), outwash terraces Retreat of glaciers from some point far down Tetsa Valley "Old" till (early Wisconsin?)	
	LAST INTERGLACIAL						
	PRE-WISCONSIN						

* Dashed lines indicate time intervals

TABLE 4.—LATE GLACIAL AND POST-GLACIAL HISTORY OF AREAS ALONG THE ALASKA HIGHWAY AND IN SURROUNDING REG

RITISH COLUMBIA		SOUTHEASTERN YUKON AND ADJACENT BRITISH COLUMBIA			SOUTHWESTERN YUKON	
mit Lake Denny)	Muncho Lake (Denny)	Liard and Rancheria (Denny)	Cassiar District (Johnston, 1926; Kerr, 1934; Holland, 1940)	Teslin Mountains (Denny)	Slims, Alsek, and Shakwak Valleys (Sticht, 1951, Table 6)	Ar Klu (John
ings on moun- summits and turf-banked ces and re- forms just timber line erraces	River terraces	River terraces		Volcanic ash Dissection of terraces Formation of beaches (?) on Teslin Lake Sand dunes	Glaciers retreat; climate becomes warmer and wetter. Deposition of Slims Valley (eolian) silt. Drainage of recent ice-dammed lake in Alsek and Shakwak Valleys near Pine Creek A marked advance of glaciers; climate becomes cooler and drier. Deposition of Slims Valley (eolian) silt. Ice-dammed lake in Alsek and Shakwak Valleys near Pine Creek Deposition of volcanic ash	Gray, s blown thin l Under inches ash. ness inches facts
ion of boul- n talus above r line					Glaciers far back in mountains. Many small glaciers of today not in existence. Long warm interval; warmer and wetter than formerly and warmer and wetter than today. Deposition of Kluane (eolian) silt becomes very slow and finally almost ceases. Development of old soil profile in Kluane silt. Dissection of gravel deposits	Buried Top 4 red b tains low al yellow lying stratu Fine s with pebble gravel
s in cirques 5800 feet), ly post Wis- read perma- and intense action. thing of sur- f "old" till d by solifluc- outh moraine s east of Sum- ake (4350 to eet), outwash ea of glaciers some point lown Tetsa till (early nain?)	Development of terrace north of Muncho Lake (2800 feet) when lake basin ice-filled Retreat of glacier from point below junction with Liard River (below 1400 feet)	Development of pitted outwash plains and of moraine and blockfield near Watson Lake (2300 feet) Retreat of glaciers from far down Liard Valley, perhaps as far as junction with Fort Nelson River, and from Rancheria and Swift valleys	Last glaciation	? Sand dunes Widespread permafrost, intense frost action. Boulder-stripes on peak north of Mount White Disappearance of ice from basins of large lakes Development of terraces in area north of big lakes, probably ice-filled Retreat of ice from Wisconsin glacial maximum	Glaciers retreat; climate becoming warmer and wetter. Deposition of Kluane (eolian) silt and of large gravel fans in Slims Valley. Ice-dammed lakes in Alsek and Shakwak Valleys near Pine Creek. Gravel fans develop in Shakwak Valley after ice-dammed lake is drained Glaciers retreat; climate colder and drier than today. Glaciers occupy all of Slims and Alsek Valleys. Ice-dammed lakes in Takhini and Dezadeash Valleys	
			? Buried channels and lime-cemented gravels			
			? Older glaciation			

e time intervals, for example the post-glacial optimum, tentatively identified in several areas. Within any one time interval, the table does not attempt

TABLE 4.—LATE GLACIAL AND POST-GLACIAL HISTORY OF AREAS ALONG THE ALASKA HIGHWAY AND IN SURROUND

COLUMBIA		SOUTHEASTERN YUKON AND ADJACENT BRITISH COLUMBIA			SOUTHWESTERN
ake)	Muncho Lake (Denny)	Liard and Rancheria (Denny)	Cassiar District (Johnston, 1926; Kerr, 1934; Holland, 1940)	Teslin Mountains (Denny)	Slims, Alsek, and Shakwak Valleys (Sticht, 1951, Table 6)
n moun- ts and banked d re- s just r line	River terraces	River terraces		Volcanic ash Dissection of terraces Formation of beaches (?) on Teslin Lake Sand dunes	Glaciers retreat; climate becomes warmer and wetter. Deposition of Slims Valley (eolian) silt. Drainage of recent ice-dammed lake in Alsek and Shakwak Valleys near Pine Creek A marked advance of glaciers; climate becomes cooler and drier. Deposition of Slims Valley (eolian) silt. Ice-dammed lake in Alsek and Shakwak Valleys near Pine Creek Deposition of volcanic ash
boul- above					Glaciers far back in mountains. Many small glaciers of today not in existence. Long warm interval; warmer and wetter than formerly and warmer and wetter than today. Deposition of Kluane (eolian) silt becomes very slow and finally almost ceases. Development of old soil profile in Kluane silt. Dissection of gravel deposits
cirques eet), t Wis- erma- ntense of sur- i" till olifluc- moraine of Sum- 350 to utwash aciers point Tetsa (early	Development of terrace north of Muncho Lake (2800 feet) when lake basin ice-filled Retreat of glacier from point below junction with Liard River (below 1400 feet)	Development of pitted outwash plains and of moraine and blockfield near Watson Lake (2300 feet) Retreat of glaciers from far down Liard Valley, perhaps as far as junction with Fort Nelson River, and from Rancheria and Swift valleys	Last glaciation	? Sand dunes Widespread permafrost, intense frost action. Boulder-stripes on peak north of Mount White Disappearance of ice from basins of large lakes Development of terraces in area north of big lakes, probably ice-filled Retreat of ice from Wisconsin glacial maximum	Glaciers retreat; climate becoming warmer and wetter. Deposition of Kluane (eolian) silt and of large gravel fans in Slims Valley. Ice-dammed lakes in Alsek and Shakwak Valleys near Pine Creek. Gravel fans develop in Shakwak Valley after ice-dammed lake is drained Glaciers retreat; climate colder and drier than today. Glaciers occupy all of Slims and Alsek Valleys. Ice-dammed lakes in Takhini and Dezadeash Valleys
			? Buried channels and lime-cemented gravels		
			? Older glaciation		

intervals, for example the post-glacial optimum, tentatively identified in several areas. Within any one time interval, the table does no

IN SURROUNDING REGIONS*

SOUTHWESTERN YUKON			CENTRAL YUKON	ALASKA	
Valleys	Artifacts, Kluane Lake (Johnson, 1946)	Wolk Creek Area; St. Elias Range (Sharp, R. P., 1942; 1951)	Carmacks District (Bostock, 1936)	Capps (1915, 1932)	Taber (1943)
armer and wet-ian) silt. Drain-Alsek and Shak- e becomes cooler ey (eolian) silt. hakwak Valleys	Gray, sandy, wind-blown soil with thin lenses of peat. Underlain by 5 inches of volcanic ash. Total thickness 10 to 15½ inches. No artifacts	Widespread permafrost (?) Modern advance of glaciers culminating between A. D. 1840 and 1890	Volcanic ash	Volcanic ash about 1400 years ago	Gradual amelioration of climate. Permafrost residual from Pleistocene
y small glaciers warm interval; nd warmer and Kluane (eolian) almost ceases. Kluane silt. Dis-	Buried soil profile. Top 4 to 6 inches red brown, contains artifacts; below about 6 inches yellow gray overlying gray substratum Fine sand and clay with a few small pebbles, overlying gravel	Post-Wisconsin xerothermic interval. No permafrost (?) Latest—and post-(?)—Wisconsin glaciation Late Wisconsin glaciation			
armer and wet-ilt and of large ammed lakes in e Creek. Gravel ter ice-dammed rier than today. ek Valleys. Ice-deash Valleys			Moraine (?) in cirque on north side of Dawson Range Maximum of last glaciation	Wisconsin glaciation	Period of refreezing Period of deep thawing and erosion. The main interglacial period. Climate warmer than at present. Removal of silt. Ice probably still present in higher mountains Period of deep freezing. Climate cooler than at present. Active frost disintegration of bedrock. Development of mountain glaciers Period of gravel accumulation. Climate arid and moderately cold. Deposition of gold-bearing gravels Late Tertiary. Uplift and dissection. Climate warmer than at present. Deep rock decay
			Older boulder clay in vicinity of Nansen Creek	Pre-Wisconsin glaciation	

e table does not attempt to correlate events from one area to another.

tion from slopes with fresh open breaks in the vegetation to slightly gentler slopes where no open breaks or recent scars are visible; nevertheless, the trees are tilted or have curved trunks, and the surface is slightly irregular. The slumping is more pronounced on south-facing slopes than on north-facing slopes. Slumping probably took place when the ground had thawed only to the base of the root zone. In mid-June 1943, most of the movement for the season apparently had ceased.

Turf-Covered Hillocks in the Summit Lake Area

Turf-covered hillocks were not seen along the Highway southeast of Whitehorse except near Summit Lake. Near the headwaters of the North Fork Tetsa River (Pl. 4, Loc. 4) the valley sides are covered by turf-covered hillocks as much as 2 feet high and several feet across. The vegetation is alpine tundra with many patches of an alpine scrub of dwarf birch and willow. Elsewhere the alpine tundra in the Summit area is likely to be hummocky wherever the ground is damp. The hummocks are less than a foot high and resemble those of upland pastures in northern New England.

Ice Ramparts

At the east end of Summit Lake is a small ice rampart about 10 feet wide and 3 to 4 feet high. A similar one, 2 to 3 feet high and several feet wide, is seen along the eastern shore of Watson Lake.

Summary of Late Glacial and Post-Glacial History

A generalized, incomplete picture of the late glacial and post-glacial history of the northern Rocky Mountains and southeastern Yukon is all that can be given here. Table 4 presents this history with little or no attempt to correlate events from one area to another.

The oldest known glacial deposits are the "old" till in the Summit Lake area. This material is probably early Wisconsin; at least it antedates the last substage of mountain glaciation. At the maximum of the last stage of mountain glaciation, glaciers extended from the Rocky Mountains eastward into the lowlands beyond the mountain front and perhaps down the Liard Valley almost to its junction with the Fort Nelson River. It is not known whether the ice completely covered the Rocky Mountains at this time. After this ice advance, lakes formed in the region east of the Rocky Mountains prior to an advance of the Laurentide (Keewatin) ice sheet, which extended approximately as far west as the mountain front. During or after the retreat of this ice sheet, other lakes (Beatton) formed in the area east of the Rocky Mountains. The advance of the Laurentide ice sheet at the late Wisconsin maximum built the Altamont moraine in Alberta. At that time, the region from Dawson Creek to Fort Nelson may have been in part ice-free, and have had a rigorous arctic climate that resulted in widespread permafrost and active congeliturbation.

During the waning of the last stage of mountain glaciation, moraines and terraces were formed near Summit and Muncho lakes, in the Liard Valley and Teslin Mountains. The large lakes of southeastern Yukon suggest a retreatal substage when the lake basins were ice-filled and the terraces to the north were built by meltwater. Moraines in cirques in the Summit Lake area may represent a very late Wisconsin substage, or possibly a latest- and post-(?)-Wisconsin substage, as suggested by R. P. Sharp (1951) in the Wolf Creek area, Yukon.

At present, formation of boulder rings and mantling of slopes with congeliturbates are not as active in some localities as they were in the past. This suggests that permafrost was widespread and congeliturbation was a dominant process in a very late glacial or early post-glacial time prior to the post-glacial optimum.

The small areas of prairie in the Beatton Valley, in a region now largely covered by black spruce (burned), suggest that prairies were once much more extensive and formed a stage in the development of the natural vegetation in the Beatton Valley (Raup and Denny, 1950). The development of these prairies probably coincided with the advent of the post-glacial optimum (Raup, 1946). The recent change from prairie to forest might be taken as evidence that the climate has become slightly colder in recent time. Raised beaches on the shore of Teslin Lake suggest a recent slight decrease in lake levels.

Relation of Late Quaternary History to Botanical and Archaeological Problems

Development of Vegetation

The origin and distribution of vegetation along the Alaska Highway[7] are considerably dependent upon late Quaternary events, particularly climatic and physiographic changes. The return of plants to glaciated regions has been conditioned by the availability of, first, land surface under climates that would permit the growth and persistence of plants and, second, populations of species with suitable habitat requirements and sufficient migratory capacity to invade the recently deglaciated regions (Raup, 1946).

Formerly it was assumed that the forests moved southward in front of the advancing ice sheets and later readvanced northward following deglaciation. Current botanical thought now favors the hypothesis of destruction of forests rather than their migration. In a recent paper on the development of vegetation in the Athabaska-Great Slave Lake region, Raup (1946) advances the theory that the coniferous forests in the northern interior plains of the continent were eliminated completely by the Wisconsin ice sheets. This region derived its post-glacial forests from eastern and western elements that persisted in the cordilleran areas and in the east.

The relative ages of the various land surfaces along the Highway are obviously important in any consideration of forest development (See Raup, 1946; 1947). The higher mountains along the Highway were probably the first to be uncovered by the ice, followed by or contemporaneous with the high plateau areas such as the higher parts of the Yukon and Alberta Plateaus and related surfaces in the mountains. Ice presumably remained longest in the valleys in the higher mountains. River terraces and flood plains are the youngest surfaces in the region. Surfaces of intermediate age are the bottoms of post-glacial lakes that remained until their ice or morainic dams disappeared. Unfortunately it is not possible at present to correlate the retreatal substages of the mountain glaciers from one area to another.

[7] The distribution of forest types is shown in Plate 3 (Raup, 1945, Figs. 1, 2; Raup and Denny, 1950, Pl. 9).

In late Wisconsin time, when glaciers were retreating into the mountains, there was a long period of active congeliturbation. The results of frost action indicate that congeliturbation was of an intensity that would have been impossible had the surfaces been covered with trees. The period of active congeliturbation was long enough to allow the almost complete elimination of the topographic forms that marked the original glacial deposits, and to produce the long cryoplanation slopes now so characteristic of the landscapes. These slopes have their most pronounced development in southwestern Yukon and Alaska, but they are also prevalent in the eastern foothills of the Rockies, in the upper Liard region, and in southeastern Yukon. With the exception of those on the more recent floodplain, lake shore, and morainic deposits, most if not all of the forests along the Alaska Highway are on surfaces whose solifluction features could not have been formed if forests had been present.

To the relative age of surfaces, therefore, must be added the problem of their relative freedom from an intensity of frost action that may have prevented the development of forests.

There is evidence that the forests of the northern great plain of the continent achieved the repopulation of that area at a time near the advent of the post-glacial optimum (Raup, 1946). If so, forests probably did not reach the upper Liard River and upper Yukon basins until the same period or later. If the forests are very young, the question arises as to what effect their development has had on geological processes, especially those related to frost action. Evidence has been presented to show that permafrost and congeliturbation have decreased in extent and in intensity in the recent past, perhaps owing in part to the encroachment of forests over the landscape. Forests keep the ground in shade and decrease daily freezing and thawing. Perhaps the processes that produced the ancient boulder rings on the outwash terrace southeast of Summit Lake, British Columbia, were stopped by the growth of the pine and spruce forest. Spruce commonly grows on frozen soil but pine does not, perhaps because the deep taproot of the pine "requires soils in which the level of permanent frost is low" (Raup, 1946; after Pulling, 1918). A forest, however, probably would act as an

insulator and preserve permafrost. More probably the development of pine on the terrace near Summit Lake indicates that an amelioration of the climate destroyed the permafrost and permitted pine growth.

Some suggestion of recent climatic changes is to be seen in the occurrence of prairies along the Highway east of the Rocky Mountains. The possibility that the flat floor of the Beatton Valley formerly was covered by prairie has been suggested by Raup (1945). Parts of the valley, not visited but covered by aerial photographs taken since 1943, show what are apparently small areas of prairie now surrounded by muskeg (Raup and Denny, 1950). These prairie relics may well be remnants of once extensive prairies that covered much of the floor of the Beatton Valley; probably they were developed in the post-glacial optimum. Their disappearance suggests a slight cooling of the climate in recent time, but there is no geological evidence in the Beatton Valley for these changes (See Raup, 1946; 1947; Porsild, 1938; 1945). The varved sediments which underlie much of the valley floor show disturbed stratification or none at all to depths of as much as 8 feet in some places. This disturbance is probably the result of frost heaving during an interval of intensive frost action prior to the post-glacial optimum. Similar varved sediments near Teslin Lake are disturbed to depths of only a foot or two.

The source areas of the trees that invaded the Highway region are still uncertain. The trees are of distinctly western American affinity (Raup, 1945), either species peculiarly western in their general distribution or western varieties of species complexes that have wide boreal American ranges. Raup believes that these species may have lived through the Pleistocene in western refugia, such as in the foothills of the central Rockies or on the western coastal strip.

That the Yukon basin in central Yukon Territory and Alaska was not glaciated in the Wisconsin has long been recognized. However, the possibility that there were large nunatak areas in the mountains to the south (presumably in southeastern Yukon and northern British Columbia) is not supported by the admittedly fragmentary geological data available (See Bostock, 1948, p. 10). Large areas remain to be explored, especially north of the Liard River. In any case, it is difficult to see how forests could have survived in such refugia through the vicissitudes of the periglacial climate of these regions, accompanied by the violent stirring of soils by frost action.

The distribution of forest types along the Highway is related in part to local variations in the kind of ground on which they grow: soil texture, drainage, slope, exposure, and altitude (Raup and Denny, 1950). A striking example of this relationship is the distribution of the black spruce on heavy clay soils derived from lake deposits or clay till. Throughout the foothills east of the Rocky Mountains and in the Liard Valley there is a close correlation between the occurrence of heavy clay soils and black spruce in a muskeglike forest where it is mixed with white spruce and lodgepole pine (Pl. 3). Another example is lodgepole pine in relatively open stands on sandy and gravelly stream terraces or outwash plains.

Archaeological Sites

Stone artifacts were found by Mr. Frederick Johnson at many places along the Highway during the field season of 1944 (Johnson, 1946). The first site at which stone tools were found is on the southwest shore of the Little Arm of Kluane Lake, and extends for about a mile along a terrace that is about 8 feet above the present high-water level of the lake. An excavation at this site is described by Johnson (1946, p. 185) as follows:

"The upper stratum varies in thickness from 10 to 15½ inches and is composed of several layers. The first is made up of some 4 inches of gray, sandy, wind-blown soil, in which are roots of various kinds. Underlying this is a layer of volcanic ash which averages more than 5 inches in thickness. At various levels between the ash and the gray sandy soil are large, thin, lenses of peat, that is, compact masses of dark-colored vegetal material bound together by fine roots.

"The lower stratum . . . is composed of a deposit of fine sandy clay in which are rare, very small pebbles. The deposit is probably of eolian origin. It was divided into three different-colored zones. The uppermost was red-brown and its thickness varied from some 4 to 6 inches. The thickness of the second, yellow-gray in color, averaged about 6 inches. The third was identified as gray, fine sand and clay; it was 4 to 6 inches thick. Below this appear gray sand and pebbles, the number and size of the pebbles increasing with depth until a yellow gravel, probably water-laid, is encountered.

"All artifacts discovered in situ were in the red-

brown layer of the lower stratum. . . . The conclusion is inescapable that the artifacts were laid down when the red-brown layer was exposed. It should be emphasized that no prehistoric artifacts or chips were found in the upper stratum."

Sticht (1951) calls the upper 4 inches of wind-blown soil, the "Slims Valley silt" and the lower stratum, the "Kluane silt." The latter he interprets as an eolian mantle, in the upper part of which was developed a red-brown soil profile that perhaps was formed during the post-glacial climatic optimum. The artifacts appear to have been dropped in the Kluane silt toward the close of the period of its disposition and before the development of the red-brown soil profile.

"The archaeological sites found east and south of Whitehorse are, at the moment, significant chiefly because of their geographic setting and because they extend the distribution of a cultural unit over a distance of about 1000 miles" (Johnson, 1946, p. 186). In this area the artifacts occur at or near the ground surface, commonly just beneath a surface layer of organic matter, and their stratigraphic position is indeterminable in relation to any deposits that might be the correlative of the ancient soil of southwestern Yukon. Artifacts were found along Little Atlin Creek north of Mount White, near Squanga Lake, near Ten Mile Creek on Teslin Lake, Teslin Village, Muncho Lake, Toad River-Muncho Lake divide, Toad River-Racing River divide, in Summit Pass, and east of the Rocky Mountains near Mill Creek and at the Buckinghorse River.

Little can be said about the relation of the artifacts found in the area southeast of Whitehorse to the glacial chronology except that they are apparently very young. If the artifacts represent the same time interval (a strictly unsupported assumption), they define the position of the post-glacial optimum in the stratigraphy of the region. The finds near Mill Creek, west of Fort Nelson, British Columbia, come from beneath a 1-foot layer of peaty material in a burned muskeg. This suggests that the muskeg of black spruce developed after the post-glacial optimum.

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DYNAMIC BASIS OF GEOMORPHOLOGY

By Arthur N. Strahler

Abstract

To place geomorphology upon sound foundations for quantitative research into fundamental principles, it is proposed that geomorphic processes be treated as gravitational or molecular shear stresses acting upon elastic, plastic, or fluid earth materials to produce the characteristic varieties of strain, or failure, that constitute weathering, erosion, transportation and deposition.

Shear stresses affecting earth materials are here divided into two major categories: gravitational and molecular. Gravitational stresses activate all downslope movements of matter, hence include all mass movements, all fluvial and glacial processes. Indirect gravitational stresses activate wave- and tide- induced currents and winds. Phenomena of gravitational shear stresses are subdivided according to behavior of rock, soil, ice, water, and air as elastic or plastic solids and viscous fluids. The order of classification is generally that of decreasing internal resistance to shear and, secondarily, of laminar to turbulent flow.

Molecular stresses are those induced by temperature changes, crystallization and melting, absorption and desiccation, or osmosis. These stresses act in random or unrelated directions with respect to gravity. Surficial creep results from combination of gravitational and molecular stresses on a slope. Chemical processes of solution and acid reaction are considered separately.

A fully dynamic approach requires analysis of geomorphic processes in terms of clearly defined open systems which tend to achieve steady states of operation and are self-regulatory to a large degree. Formulation of mathematical models, both by rational deduction and empirical analysis of observational data, to relate energy, mass, and time is the ultimate goal of the dynamic approach.

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Introduction

The aim of this paper is to outline a system of geomorphology grounded in basic principles of mechanics and fluid dynamics, that will enable geomorphic processes to be treated as manifestations of various types of shear stresses, both gravitational and molecular, acting upon any type of earth material to produce the varieties of strain, or failure, which we recognize as the manifold processes of weathering, erosion, transportation and deposition. The concepts set forth in this paper have been established as guiding principles underlying the quantitative investigation of erosional landforms by the writer and his associates under Contract N6 ONR 271, Task Order 30, Project No. NR. 089-042, Office of Naval Research, Geography Branch. The writer is grateful to Professor W. C. Krumbein of Northwestern University and Professors Sidney Paige and Donald Burmister of Columbia University for their kindness in

reading the manuscript and suggesting various improvements, and to Mr. Samuel Katz of the Lamont Geological Observatory for developing the mathematical analysis of river profiles.

For more than half a century, the study of landforms in North America was dominated by an explanatory-descriptive method of study used by W. M. Davis and his students. Davis himself maintained that the aims of his method were geographic; that the consideration of process was introduced merely to permit an orderly genetic system of landform classification. The weakness in understanding of geomorphic processes (and hence also a weakness in the understanding of the origin of landforms) has not been confined to the American continent. The geomorphologists of France and England, closely attached to schools and departments of geography, have also tended to give much attention to descriptive, deductive studies of landform development and to regional geomorphological treatments. Even in so distinctively different a treatment as Walther Penck's morphological analysis, processes and forms are analyzed by a deductive method safely removed from the reality of existing landforms and without mention of basic principles of soil mechanics and fluid dynamics.

If geomorphology is to achieve full stature as a branch of geology operating upon the frontier of research into fundamental principles and laws of earth science, it must turn to the physical and engineering sciences and mathematics for vitality which it now lacks. The geomorphic processes that we observe are, after all, basically the various forms of shear, or failure, of materials which may be classified as fluid, plastic, or elastic substances, responding to stresses which are most commonly gravitational, but may also be molecular.

Unless the fundamental nature of materials is understood, we are in a poor position to add anything worthwhile to what is already largely self-evident concerning the behavior of streams, landslides, glaciers, or wave-induced currents. We cannot hope for anything better than a superficial knowledge of the form and motion of a sand dune unless we can interpret the dune in terms of aerodynamic principles; or of a stream profile unless we understand the principles of fluid dynamics and the transportation of sediment; or of the moulding of a drumlin unless we study the flowage and fracturing of ice; or of the production of an off-shore bar unless we know something of the dynamics of waves and the transport of sediment by oscillating or pulsating wave-induced currents; or of the causes of a great earth-flow unless we can appreciate the principles of plastic flowage.

To delegate to the civil engineer all fundamental research on geomorphic processes and forms has certain disadvantages. With his attention focused upon problems dealing with man-imposed modifications of the natural landscape, the engineer may have neither the time nor the inclination to investigate a broad range of natural phenomena where they are best displayed. Furthermore he is likely to have only a limited acquaintance with geologic materials and forms, whereas the geomorphologist, trained as a geologist, has built up a life-time store of information and experience, much of it relating to theoretical and historical aspects of geology.

Although the study of fundamental principles of geomorphology by engineers is to be welcomed and encouraged, there is a real danger that the engineer will find it necessary to take over an increasingly greater proportion of geomorphic research and thereby cut it off from its most logical parent, the field of theoretical geology. A specific example is the field of landslides and related gravity movements. The geomorphologists now owe their most penetrating analysis of the fundamental principles of these phenomena to research by specialists in soil physics and soil mechanics. Karl Terzaghi's (1950) work is outstanding in this respect. Far from being a supplier of basic theoretical knowledge to civil engineers in this field, as he should be, the geomorphologist has been receiving this information from them.

It is appropriate in this paper, which is philosophical in nature, to clarify the function of time in geomorphology. Two quite different viewpoints are used in dynamic (analytical) geomorphology and in historical (regional) geomorphology. The student of processes and forms *per se* is continually asking "What happens?"; the historical student keeps raising the question "What happened?". Bucher (1941) has aptly labeled the two types of geological information as *timeless* and *timebound* knowledge respectively. It is largely with the timeless knowl-

edge that the field of dynamic geomorphology deals. The principles are usually most easily discovered by a study of contemporary processes and existing forms, but the dynamic geomorphologist will refer to any part of the geologic record for evidence which will increase his understanding. He is not, however, primarily concerned with the actual series of events within a particular geologic period and in a particular geographical location, as is the historical geomorphologist. Although many historical-regional geomorphic investigations have been ably conducted with minimum reference to geomorphic processes, it is only reasonable to suppose that a better knowledge of how processes operate and normal forms evolve will increase the effectiveness of historical studies and reduce the likelihood of drawing erroneous inferences of past events.

On the other hand, studies of dynamic geomorphology based on existing landforms cannot be prosecuted in ignorance or disregard of past changes of climate, and hence, of relative rates and importance of various processes dependent on climate, during the long period required to develop the forms. One cannot, for example, extrapolate rainfall intensity-frequency-duration statistics of the past 50 years back over a period of 50,000 years in correlating drainage basin forms with rainfall characteristics. The dynamic geomorphologist must be alert to evidences of important differences in processes operating during earlier stages of development of the forms he is studying. Thus historical-regional geomorphic treatment cannot be divorced from dynamic investigations. The difference in the two types of study lies in proportion of each one involved, for neither can successfully be pursued independently of the other.

Basis of a Dynamic Approach

Type of Stress

Table 1 outlines the organization of a dynamic treatment of geomorphology. The first basic subdivision is made according to the nature of the stresses involved: (A) *gravitational stresses* and (B) *molecular stresses*. The gravitational stresses act upon all earth materials. Where these have a sloping surface, components of the gravitational stress tend to produce movement of matter downslope. Under this heading, therefore, are placed both the mass gravity movements and the erosional-transportational fluids: water, ice, and air. Energy transformed during the operation of these processes is for the most part the potential energy of position or elevation. Earth materials raised by orogenic or epeirogenic crustal movements are gradually moved to lower elevations, with an accompanying transformation of potential energy to kinetic energy of heat or motion. Water raised from sea level to divide regions by expenditure of solar energy in atmospheric heating and turbulence makes its way to lower elevations, at the same time expending a small fraction of the total transformed energy in overcoming the cohesion or bonding of rock and soil particles and in transporting them to lower levels.

The work of wind is a somewhat different form of response to gravitational stress in that solar heating has set up differences in air mass densities which tend to be equalized by air flow from regions of higher to lower pressure. Some winds (katabatic) are actually a form of downslope flowage and are thus akin to streams and glaciers. Shore currents may be wave-induced, in which case the generation of waves may be traced back to gravitational stresses through the winds which produced them, or tide-induced, in which case gravity flow responding to differences in water level occurs. The general arrangement of materials in Table 1A is in order of decreasing resistance to shear stresses.

The molecular stresses (Table 1B) may act in any direction with respect to gravity, and in a homogeneous or isotropic soil or rock material may be distributed at random in all possible directions through a given point. The characteristic movement is that of dilatation of a mass. Downslope movement is not essential, but creep to lower levels is inevitable on any slope, because a component of the gravitational stress then adds its vector to the otherwise random stress distributions.

Table 1C lists chemical processes, which are not themselves stress-producing, although of great importance in geomorphology. Listed here are chemical reactions and simple solution. Such chemical processes as hydrolysis and oxidation, which would produce expansive stresses

by increasing volume and developing adsorptive properties, have been placed under molecular stresses.

Type of Material

Three fundamental types of materials can be recognized, although the distinction in a given

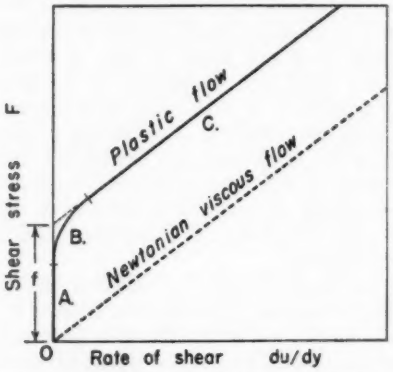

FIGURE 1.—RELATION OF RATE OF SHEAR TO SHEAR STRESS IN PLASTIC SOLIDS AND VISCOUS FLUIDS

case may be difficult to make: (1) rigid, or elastic, solids, (2) plastic solids, and (3) fluids.

An elastic solid is a mass of rock or soil which possesses elasticity and yields by elastic strain. Ideally this strain follows Hooke's Law, in which strain is proportional to stress. Stresses of greater magnitude cause failure by rupture along discrete planes and may take the form of sliding movement on shear planes or pulling apart on widening tension fractures. We will assume for the moment that there is little or no region of slow, permanent deformation (creep) between the development of elastic strain and rupture. Our definition of elastic solid is here broadened to include homogeneous soil or incoherent bedrock, which, in a densely packed state without excessive moisture, may act as an elastic continuum (Krynine, 1947, p. 92).

A plastic solid is a material that deforms by distributed intermolecular or intergranular shear—that undergoes flowage in the manner of a fluid, if the shear stresses exceed a limit set by the internal cohesion or friction of the mass. A law of plastic deformation may be stated as follows: (Burmister, 1948, p. 93)

$$\frac{du}{dy} = \frac{(F - f)}{\eta}$$

where du/dy = rate of shear (rate of change of velocity, u, with respect to depth, y)
F = shear stress
f = yield limit
η = consistency of the material.

Figure 1 illustrates the principles of plastic flow. In region A, increasing shear stress is within the limits of internal resistance of the material and no shear occurs. In region B, flowage of slow but increasingly rapid rate sets in. This is the region of plastic creep. In region C, flowage is essentially that of a true fluid in that rate of shear is directly proportional to applied stress. Bingham (1922) has described a plastic material as one which "has a property of permanently supporting a shearing stress less than a certain critical yield value with only slight total deformation, but for greater shearing stresses plastic flow takes place at a continuous uniform rate, as for a highly viscous fluid."

A fluid is a substance that offers little resistance to change in form and is incapable of any internal adjustment that will enable it to maintain equilibrium at rest while subjected to shear stress, however small (Dodge and Thompson, 1937, p. 1). Fluids include both gases and liquids, hence the study of geomorphic processes involving flow of water or wind is governed by principles of fluid flow. Whereas liquids are largely incompressible and deform by changing shape, gases change readily both in form and in volume.

A perfect fluid, one which offers no resistance to deformation, does not exist. All fluids possess viscosity, a quality of resisting shear stress. Energy is required to overcome this resistance, hence stress must be applied continually to maintain shear in a fluid, and the rate of shear is proportional to the applied stress, as stated by the following equation:

$$\frac{du}{dy} = \frac{F}{u}$$

Tab

A. Gravitational Stresse

	MATERIAL INVOLVED IN MOTION	PROPERTIES OF MATERIAL	TYPE OF FAILURE (STRAIN)
1.	Crystalline rocks, arenaceous rocks, limestones; dry soils	Rigid, elastic solid or "elastic continuum" Obeys Hooke's Law: Strain ∝ stress	Sudden rupture along shear surfaces or tension fractures.
2.	Glacial Ice, near surface.	Plastic solid in region of elastic behavior.	Sudden rupture along shear surfaces or tension fractures.
3.	Argillaceous rocks and soils	Plastic solid in region of slow creep.	Continuous, slow laminar flow (distributed shear)
4.	Unconsolidated rock, soil + water	Plastic solid in region of flowage. Obeys Bingham's law: Rate of shear proportional to stress, above a yield limit.	Plastic flow (shear between grains) when stress exceeds yield limit. Movement ceases below yield value. Flow laminar or turbulent.
5.	Glacial ice under heavy load.	Plastic solid with high yield value. Non-linear increase of shear rate with increase of stress.	Laminar plastic flow above yield limit. Below yield limit, returns to brittle elastic solid.
6.	Water film on sloping surface.	Newtonian fluid. No yield value. Linear increase of shear rate with stress. Subject to capillary influences.	Laminar flow, ceasing when water thins below capillary control limit.
7.	Water in permeable rock, soil. (No surface slope)	Newtonian fluid, subject to capillary influences.	a). Silts: Laminar flow following Darcy's Law. b). Sands: Mixed laminar and turbulent flow. c). Gravels: Turbulent flow.
8.	Water layer on sloping surface	Newtonian fluid.	Sheet runoff in turbulent or mixed turbulent-laminar flow.
9.	Water in sloping linear channel	Newtonian fluid.	Stream flow, turbulent except in bed layer.
	Indirect Responses to Gravitational Stresses		
10.	Standing water bodies. Oceans, lakes.	Newtonian fluid.	Turbulent flow as a). Pulsating or oscillating currents caused by waves. b). Tide-induced currents.
11.	Air	A gas: compressible fluid of extremely low viscosity.	Turbulent flow induced by pressure gradients. (gravitational stress on air masses)

Table 1. DYNAMIC BASIS OF GEOMORPHOLOGY

Stresses

AILURE N)	GEOMORPHIC PROCESSES AND FORMS
along or ures.	Landslides: Slump, slide (compressional stress); rock-fall (tensional stress).
e along or es.	Crevassing, overthrusting, calving of glaciers.
w laminar ed shear)	Large-scale creep phenomena. Superficial deformation of clays.
ar between s exceeds ement ld value. urbulent.	Solifluction, earth flow, mudflow, highly turbid streams, turbidity currents.
flow t. Below urns to id.	Continental and alpine glaciers. Erosion forms due to ice abrasion. Depositional forms moulded by ice flow.
ceasing s below limit.	Sheet runoff on slopes and rock surfaces. Slope reduction by removal of ions, colloids, clays. Fluting, grooving of limestone.
r flow 's Law. laminar flow. ent flow.	Infiltration of precipitation, carrying down of ions, colloids, clays, silts (illuviation). General slope reduction. Karstic forms in highly soluble rocks.
turbulent nt-laminar	Slope erosion, transportation. Slope forms of fluvial drainage basins.
rbulent ayer.	Stream erosion, transportation, deposition. Drainage systems. All fluvial landforms.

Stresses

as scillating by waves. urrents.	Shoreline processes of erosion, transportation, deposition. Shoreline landforms: cliffs, benches, beaches, bars, spits.
induced dients. on air masses)	Wind erosion, transportation, deposition. Deflational and abrasional forms. Dunes, loess.

B. Molecu

	MATERIALS INVOLVED	PROPERTIES OF MATERIAL	STRESS AND CAUS
1.	Rock: strong, hard crystalline, glassy or crystal aggregate.	a). Elastic solid, non-homogeneous.	Shear stress due to non-uniform expansio contraction in cyclic temperature changes
		b). Elastic solid, homogeneous	Shear stress set up by thermal gradient from surface heating.
2.	a). Permeable rock + water b). Clay soils	a). Elastic solid. b). Plastic solid.	a). Shear stress set up b interstitial ice crys growth. b). Stress from growth o ice lenses, wedges.
3.	Permeable rock or soil + water and salts	Elastic solid or elastic continuum	Shear stress set up b interstitial growth of salt crystals.
4.	Rock or soil + colloids and water	Elastic or plastic solid	Shear stress set up dilitation accompanyi water adsorption and dr
5.	Rock or soil + capillary water	Elastic or plastic solid	Shear stress set up by dilitation accompanyin changes in capillary film tension.
6.	Rock or soil + plant roots	Elastic or plastic solid	Shear stress set up by swelling of rootlets und osmotic pressure.
7.	Strong, hard monolithic bedrock	Elastic solid	Shear stresses of tecto origin stored as elastic strain at depth.

C. Chem

	MATERIALS	PROCESS
1.	Soil, rock + acids, water	Reaction between acid ions and mineral surfaces. Removal of products in solution.
2.	Soil, rock + water	Simple solution (ionization) of susceptible minerals.

Table 1. DYNAMIC BASIS OF GEOMORPHO

A. Gravitational Stresses

PROPERTIES OF MATERIAL	TYPE OF FAILURE (STRAIN)	GEOMORPHIC PROCESSES AND FORMS
Rigid, elastic solid or "elastic continuum" Obeys Hooke's Law: Strain ∝ stress	Sudden rupture along shear surfaces or tension fractures.	Landslides: Slump, slide (compressional stress); rock-fall (tensional stress).
Plastic solid in region of elastic behavior.	Sudden rupture along shear surfaces or tension fractures.	Crevassing, overthrusting, calving of glaciers.
Plastic solid in region of slow creep.	Continuous, slow laminar flow (distributed shear)	Large-scale creep phenomena. Superficial deformation of clays.
Plastic solid in region of flowage. Obeys Bingham's law: Rate of shear proportional to stress, above a yield limit.	Plastic flow (shear between grains) when stress exceeds yield limit. Movement ceases below yield value. Flow laminar or turbulent.	Solifluction, earth flow, mudflow, highly turbid streams, turbidity currents.
Plastic solid with high yield value. Non-linear increase of shear rate with increase of stress.	Laminar plastic flow above yield limit. Below yield limit, returns to brittle elastic solid.	Continental and alpine glaciers. Erosion forms due to ice abrasion. Depositional forms moulded by ice flow.
Newtonian fluid. No yield value. Linear increase of shear rate with stress. Subject to capillary influences.	Laminar flow, ceasing when water thins below capillary control limit.	Sheet runoff on slopes and rock surfaces. Slope reduction by removal of ions, colloids, clays. Fluting, grooving of limestone.
Newtonian fluid, subject to capillary influences.	a). Silts: Laminar flow following Darcy's Law. b). Sands: Mixed laminar and turbulent flow. c). Gravels: Turbulent flow.	Infiltration of precipitation, carrying down of ions, colloids, clays, silts (illuviation). General slope reduction. Karstic forms in highly soluble rocks.
Newtonian fluid.	Sheet runoff in turbulent or mixed turbulent-laminar flow.	Slope erosion, transportation. Slope forms of fluvial drainage basins.
Newtonian fluid.	Stream flow, turbulent except in bed layer.	Stream erosion, transportation, deposition. Drainage systems. All fluvial landforms.

Indirect Responses to Gravitational Stresses

Newtonian fluid.	Turbulent flow as a). Pulsating or oscillating currents caused by waves. b). Tide-induced currents.	Shoreline processes of erosion, transportation, deposition. Shoreline landforms: cliffs, benches, beaches, bars, spits.
A gas: compressible fluid of extremely low viscosity.	Turbulent flow induced by pressure gradients. (gravitational stress on air masses)	Wind erosion, transportation, deposition. Deflational and abrasional forms. Dunes, loess.

	MATERIALS INVOLVED	PROPERT OF MATE
1.	Rock: strong, hard crystalline, glassy or crystal aggregate.	a). Elastic s non-ho ganeo b). Elastic s homogen
2.	a). Permeable rock + water b). Clay soils	a). Elastic s b). Plastic s
3.	Permeable rock or soil + water and salts	Elastic so or elasti continuu
4.	Rock or soil + colloids and water	Elastic or plastic so
5.	Rock or soil + capillary water	Elastic or plastic so
6.	Rock or soil + plant roots	Elastic o plastic soli
7.	Strong, hard monolithic bedrock	Elastic so

	MATERIALS	
1.	Soil, rock + acids, water	Re io Re so
2.	Soil, rock + water	Si o

ORPHOLOGY

B. Molecular Stresses

PROPERTIES OF MATERIAL	STRESS AND CAUSE	KIND OF FAILURE	WEATHERING PROCESS AND FORM
a). Elastic solid, non-homogeneous.	Shear stress due to non-uniform expansion-contraction in cyclic temperature changes.	Rupture by shear or tension fractures between grains, along cleavages, joints, bedding planes.	Granular or blocky disintegration of rocks, esp. coarse-grained crystalline rocks.
b). Elastic solid, homogeneous	Shear stress set up by thermal gradient from surface heating.	Rupture between layers paralleling rock surface.	Exfoliation of rock by fire, lightning; solar or atmospheric heating-cooling.
a). Elastic solid. b). Plastic solid.	a). Shear stress set up by interstitial ice crystal growth. b). Stress from growth of ice lenses, wedges.	a). Rupture between grains, cleavage pieces, joint blocks, beds. b). Plastic deformation of clays adjacent to ice.	a). Frost disintegration of rocks. Felsenmeer. b). Heaving of clay soils, frost mounds, polygons.
Elastic solid or elastic continuum	Shear stress set up by interstitial growth of salt crystals.	Rupture between grains, cleavage pieces, joint blocks or beds.	Effloresence, granular disintegration in dry climates. Caliche heaving.
Elastic or plastic solid	Shear stress set up by dilitation accompanying water adsorption and drying.	a). Rupture between grains. b). Plastic deformation of clays during swelling.	Exfoliation of basaltic, granitic rock upon alteration of silicates. Slaking of shales, argillaceous ss.
Elastic or plastic solid	Shear stress set up by dilitation accompanying changes in capillary film tension.	Rupture between grains or masses of soil.	Disintegration of granular permeable rocks. Heaving or subsidence of clays, silts.
Elastic or plastic solid	Shear stress set up by swelling of rootlets under osmotic pressure.	Rupture between grains, cleavage pieces, joint blocks or beds.	Disintegration of rock by prying of roots. Deformation of soils.
Elastic solid	Shear stresses of tectonic origin stored as elastic strain at depth.	Rupture of rock on planes paralleling surfaces after release of confining pressure.	Exfoliation of domes, slabs, shells. Quarry rupture, rock-burst.

C. Chemical Processes

LS	PROCESS	FORMS PRODUCED
	Reaction between acid ions and mineral surfaces. Removal of products in solution.	Lowering of rock and soil surfaces. Pitting, cavitation of rocks, esp. carbonates. Cavern and karst forms (see Table A, no. 7). Weakening of bonds between mineral grains.
	Simple solution (ionization) of susceptible minerals.	Cavitation of soluble salt formations. Slow attrition of exposed mineral surfaces.

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where y is depth above bottom
u is velocity
F is shear stress
μ is viscosity.

Under this type of flow, termed Newtonian viscous flow, the relationship of shear stress to rate of shear is linear, the slope of the straight line being proportional to viscosity. Note in Figure 1 that the line passes through the origin, signifying that no matter how small the applied stress some deformation by shear will occur.

Type of Strain

The type of failure, or strain, that may be expected as a result of the application of gravitational or molecular stresses on elastic, plastic, or fluid earth materials determines the geomorphic process and form. To some extent types of failure have already been noted in defining the fundamental properties of the materials.

Rock or soil behaving as an elastic solid fails by one of two possible forms: (1) A shear fracture may develop as a slip plane formed in the zone where shear stresses first exceed resistance to shear. Differential movement of one mass or block over another by sliding along the shear surface constitutes the type of movement and is characteristic of slump types of landslides or of the surface zone of glaciers where overthrusting occurs. (2) A widening tension fracture may develop in material subjected to tensile stresses tending to produce elongation of the mass. Examples are particularly striking on glaciers and landslides where gaping crevasses and fissures are formed. Where molecular stresses are active in a rock or soil having elastic properties, the shear fractures and tension fractures may have the extremely small dimensions of grain contacts or cleavage planes. Thus weathering phenomena involving disintegration of bedrock are largely of this type of failure.

In fluids, and in plastic solids above the yield limits, flowage occurs in two forms: laminar flow and turbulent flow, both of greatest importance in geomorphic processes. In laminar flow, parallel layers of the fluid are in motion over one another without the existence of cross currents of motion. Each layer of the fluid moves at a higher velocity than the layer adjacent to it on the one side but lower than the layer on the other side. The intermolecular forces are strong enough to prevent the development of cross-threads of particle movement.

Laminar flow can be maintained in deep water only at very low velocities. Thus even extremely slow flow in sluggish streams and tidal currents may be turbulent. Erosion and transportation of coarse-grained materials cannot occur in laminar flow; instead the settlement of suspended sediment is permitted. Somewhat more rapid flow of laminar nature is possible in thin films of water, such as in sheetflow over soil or rock surfaces or while infiltration is occurring through permeable rock or soil. The removal of ions, colloids, extremely fine clays, or silts may occur by this process and it is therefore of considerable geomorphic importance in denudation.

Laminar flow is possible at higher velocities in fluids of high viscosity than in those of low viscosity; glacier ice, which behaves essentially as a fluid above its elastic limit, flows exclusively by laminar principles. Air, by contrast, has such low viscosity that for all practical purposes any air currents of sufficient velocity to move fine sand or dust can be considered as in turbulent flow.

Turbulent flow sets in when flow velocity exceeds a critical velocity determined by factors of depth, viscosity, and roughness of the bounding surface (Hjulström, 1935, p. 238). Turbulent flow is characterized by components of velocity in many directions within the moving fluid. These have been thought of as innumerable cylindroids of rotation continually forming and dissolving in the fluid. Although an average forward velocity prevails at any one particular point in the stream of flow, the instantaneous velocities are distributed at random, both in direction and magnitude (Einstein, 1950, p. 13). The result is vertical components of flow which can sustain solid particles; hence the transportation of rock materials in suspension is possible by flowing air and water at the velocities normally prevailing in streams, wave-induced currents, and winds.

The work of running water, waves, and wind cannot be understood without a thorough appreciation of the principles of laminar and

turbulent flow. Fluid dynamics is, therefore, a cornerstone of geomorphology.

In the plastic earth materials, such as earthflows and mudflows, our definition of "flow" is broadened to include shear more or less uniformly distributed throughout the mass and occurring by the rotation or slippage of grains with respect to one another. As the particle size involved diminishes, the deformation simulates more and more perfectly fluid flow in which shear is among molecules of the liquid or gas. On the other extreme is the flow of a great landslide mass in which huge boulders of rock are incorporated. Considered in detail, such movement would not follow the laws of fluid flow, but if we think of the mass as a whole and if the landslide is very large in relation to the particles of which it is composed, the principles will apply. In any case, the type of deformation is distinctly different from the sliding movements in which a large mass moves as a unit on a single shear surface.

Before leaving the topic of general principles of types of failure in response to shear stresses, a word of comment upon the composite behavior of certain earth materials is in order. Ice and some types of argillaceous sediments respond differently to stresses depending upon the duration of the stress. A weak but continuously applied stress will cause continuous deformation (creep) by flowage, whereas a severe stress applied suddenly will be met by an elastic response.

Gravitational Stress Phenomena

Table 1A outlines the various gravitational stress phenomena, listed in a general way according to the property of the material involved. Beginning with rock, soil, or ice behaving as elastic solids, the table proceeds through the plastic earth materials to the fluids, taken in order from higher to lower viscosity—first water, then air—with laminar flow being placed ahead of turbulent flow.

Most of the processes covered by this table involve the downslope movement of material. Here gravitational stress tending to produce shear is some fraction of the total gravitational force. The slope function, which is the sine of the angle of slope, determines the intensity of erosional and transportational activities. In the case of water percolating through a permeable rock or soil, the surface slope function is absent, as the process occurs in a subsurface location. Subsurface flow lines are functions of permeability and excess hydrostatic head and may follow various directions and velocities. This distinction is worth noting because, in regions of very low surface slopes, the removal of mineral matter by percolation of ground water might achieve quantitative importance equal to that of surface runoff and mass-movement processes.

A brief explanatory statement follows for each tier of Table 1A with mention of a few representative published works concerning each:

1. Rock and soil which is relatively dry. This material possesses a high degree of internal friction or cohesion but lacks a tendency to plastic deformation; it behaves under near-surface conditions of temperature and confining pressure as an elastic solid or elastic continuum, hence yields slightly by elastic deformation until the limit of internal cohesive strength or intergranular friction is exceeded. Because this occurs along a localized surface of maximum shear stresses, the rupture is in the form of a sliding plane, as in slump blocks of bedrock or soil where an up-concave surface of sliding is formed and the mass rotates to a position where forces tending to produce shear are more nearly balanced throughout the shear zone. In other cases, particularly where a plastic substratum is in flowage, the rigid upper part of a landslide mass ruptures in nearly vertical tension fractures. The break off of overhanging rock masses on a cliff is a form of rupture of an elastic solid, but the plane of breakage is usually predetermined by joint or bedding planes along which the cohesion is much less than in the solid rock.

Yielding of elastic rock materials by rupture is characteristically followed by rapid motion of the mass, often accelerating at the rate of a freely falling body. This is because the gravitational shear stress built up before rupture is vastly greater than the stress required to move the detached body once the bonds along the fracture surface have been broken. (For further discussion of dynamics of slumping and landsliding *see* Terzaghi, 1943, 1950; Terzaghi and Peck, 1948; Krynine, 1947; and Hubbert, 1951.) Engineers have long been familiar with the

principles of stress distribution in slumps but geomorphic studies of naturally occurring slumps have not thus far used the quantitative-dynamic approach.

2. Glacial ice in the surface zone where, to all visible evidences, the behavior is that of an elastic solid. Perutz (1950) has estimated, from deformation of a pipe embedded in a glacier, that to a depth of about 50 meters no plastic flowage is observable. Crevassing, the tensional type of rupture associated with surface ice, is approximately limited to this depth. Rupture by overthrust faulting occurs in the ablational zone of a glacier where ice thickness is less than the limit required for flowage, but where stresses transmitted from points upstream exceed the elastic limit. Calving, the break-off of ice blocks from the front of a tide-water glacier, may be considered as rupture resulting from undercutting by melting or by differential vertical stresses caused by rise and fall of tide level.

3. Argillaceous rocks and clays subject to exceedingly slow, continuous creep in a downslope direction (Terzaghi, 1950; Hollingworth *et al.*, 1944). These materials are characterized by a high cohesion of particle surfaces but a relatively low coefficient of friction among the particles. A coarse sand or gravel, with no cohesion but high internal friction, is totally lacking in such plastic properties. Valleyward flowage results in development of superficial fold and thrust structures resembling large-scale tectonic structures. Thrust sheets moving long distances under gravitational stress have been postulated by geologists, and these occurrences may be thought of as intermediate in scale between tectonic and surficial forms. The materials involved in continuous slow creep would normally rupture as elastic solids under severe, suddenly-applied stresses.

4. Unconsolidated rock or soil with liberal amounts of water, such that the material behaves as a plastic solid. When shear stress exceeds the yield limit, flowage sets in at a rate governed by the intensity of the shear stress and the consistency of the material. Flowage consists of intergranular shear distributed more or less uniformly. Geomorphic phenomena included in this type of flowage are solifluction, earthflow, and mudflow, in order of decreasing consistency and increasing velocity. To this list might be added turbidity currents which, through the presence of suspended sediment, may perhaps be more exactly classed as plastic flow than fluid flow.

A definitive feature of the plastic flow phenomena is the cessation of movement while the material is still on a valley-side slope or in a sloping channel. Even though gravitational shear stress is till acting, the material may cease to move because either the stress has fallen below the yield limit (because of the mass reaching a lower slope) or the yield limit has increased (by desiccation of the mass) above the previous yield limit. Unlike a fluid, which will continue to flow until a water-level surface is attained, the plastic solid can remain stable on a slope. Plastic flowage in soils has been discussed by Burmister (1948, p. 91–101), but the principles have not yet been fully applied to the study of naturally occurring flowage movements on slopes.

5. Glacial ice under load. Characterization of glacier movement as plastic flow is based upon the recognition of a surface zone, some 30 to 50 meters thick, in which no flowage is observed in the ice, but rather rupture along tension fractures or thrust planes. Perutz (1950) calculated from observations of deformation of a steel tube sunk into the Jungfraufirn that the rigid zone extended to a depth of 50 meters, at which level a yield stress of 0.1 kg. per cm.^{2} existed. Observable deformation set in from this point downward. Rate of increase of shear with respect to shear stress was not, however, Newtonian linear increase, but increased more rapidly with depth. It is, of course, possible that slight flowage does occur in ice of less thickness, but it must be so slow as to be unimportant in relation to the time span of the existence of the ice in the surface zone where ablation is in progress.

Glacier flow is laminar because of its extremely high viscosity (consistency) and great depth. Treated in this light, glaciers would be judged incapable of lifting debris through turbulence, but must obtain their englacial load from overthrusting or from the surface. If glacier ice is regarded as a plastic solid, we have an explanation of the absence of flowage in embryonic or relict glaciers, even though they rest on a considerable slope.

Landforms included under the category of plastic glacial ice are, of course, the glaciers

themselves, both alpine and continental, and all erosional and accumulational landforms produced by the flowage of ice over ground surfaces. *Rôches moutonnées*, grooves, striae, troughs and other ice-carved forms in bedrock are included, as are drumlins and other till forms moulded of debris by ice flowage. Further progress in understanding the origin of these features must be based upon fundamental principles of ice flowage, now being studied intensively by physicists.

6. Flow of thin films of water on sloping surfaces. The water is treated as a Newtonian fluid without a yield value but controlled in its behavior by capillary forces. Water heavily charged with fine matter in suspension might be more accurately treated as a plastic solid of very low consistency. Water in thin films on soil or rock surfaces moves largely or entirely by laminar flow. This is likely to occur on smooth rock surfaces where the water is held by capillarity against the surface; on most rough soil surfaces, flow is normally mixed laminar and turbulent (Horton, 1945, p. 312).

The geomorphic activity of laminar water films is removal of ions, colloids, and fine clay particles from slopes. Fluting and grooving of limestones and basalts are the striking forms which may be attributed in part to this process, but unseen increments of slope reduction by this process may be universal in distribution and quantitatively of great importance.

7. Downward percolation of water through permeable soil or bedrock (Burmister, 1948, p. 111–115). In fine-grained materials of silt sizes, the flow is entirely laminar and follows Darcy's law:

$$v = K\frac{H}{L}$$

where v = average velocity of flow
K = a constant
H = head
L = length of soil column.

In sands, flow may be mixed laminar and turbulent; in gravels with large void spaces, flow is turbulent, above a critical velocity. Because of great variations in the cross-section of the paths of flow, the velocity is unsteady, alternating between laminar flow in the constricted passages between grains and turbulent flow in the wider voids.

The geomorphic importance of percolation may be considerably greater than indicated by textbooks and published geomorphic papers, which give this phenomenon little or no mention. Depending upon the permeability of the soil or rock, ions, colloids, clays, or fine silts may be carried down from the surface layer to lower depths or to stream channels. Accompanying compaction under direct gravitational stress lowers the ground surface. Where rainfall is heavy, vegetative cover strong, and slopes low, this process could be the major agent of slope reduction. Its relative importance will increase as the surface flow of runoff decreases in erosional intensity. Where dominant, percolation removal would be expected to modify or destroy normal fluvial-erosional topography, producing an irregular surface as in many karstic terrains.

8. Water flowing on sloping soil or rock surfaces where depth and velocity are such that turbulent flow results. The hydrodynamic characteristics of surface runoff have been studied intensively by Horton (1945), Little (1940), Schiff and Yoder (1941), and others engaged in soil investigations. Flow is mixed laminar and turbulent in some cases where depth is slight and vegetative cover creates constrictions in flow paths, but with discharges typical of torrential rainfall the flow is wholly turbulent.

The geomorphic importance of surface runoff is paramount. The development of drainage basin slopes is largely attributable to erosion and transportation by sheet flow. Because a very large percentage of the earth's landsurface is composed of fluvial erosional topography, this process may perhaps be the most important single process in forming the landscape. Geomorphologists have neglected the intensive, quantitative-dynamic study of slope development by surface runoff. This represents a glaring deficiency in a field of earth science where even adequate understanding of any process or form is a rarity.

9. Water flowing in a linear channel, i.e. as a stream. All stream flow of sufficient velocity to transport sediment and perform significant morphological work may be regarded as turbulent. The development of drainage systems,

valleys, and all of the depositional features associated with floodplains, deltas, terraces, alluvial fans, and glacial meltwaters are included in the work of streams.

Fortunately streams have been studied from the dynamic-quantitative standpoint by competent scientists over a long period of years. Among the important geological contributors are G. K. Gilbert (1914), W. W. Rubey (1933; 1938), John Leighly (1932; 1934), and Filip Hjulström (1935). Fundamental work on dynamics of streams with particular reference to the transportation of bed load and suspended load has been published in recent papers by G. H. Matthes (1941; 1949), H. A. Einstein (1950), Samuel Shulits, (1936; 1941), V. A. Vanoni (1947; 1948), Hunter Rouse (1939), A. A. Kalinske (1947), and J. F. Friedkin (1945), to list but a few examples. Future geomorphic studies of stream-formed features and of the Pleistocene and earlier history of river systems under the influences of changing conditions of discharge, load, gradient, and base level must be based upon sound principles of stream action as they are being discovered today through the experimental and quantitative field studies of the hydraulic engineer.

10. Shore-line processes, which we may regard as water currents or pulsations generated by the impingement of water waves upon a shoaling bottom or by differences of level accompanying tides. Wind-generated waves cannot be regarded as direct responses to gravitational shear stresses acting on a slope, but the winds from which the energy is derived by shear of air over water owe their existence to differences in air density. One might, of course, make a separate stress group out of propagated water waves (tsunami) generated by crustal movements. But the pure deep-water wave does not directly accomplish significant erosion or transportation at the shore line. Only when the energy of orbital motion is transformed into mass movements of water such as the swash (uprush), backwash (backrush), or longshore current, does the wave become a major agent. In these cases, gravitational stress due to differences in water level or to slope of the beach is important in producing or modifying the flow. The backwash is simply the return gravitational flow, while some longshore currents may be hydraulically induced by the raising of water level near shore by shoreward drift of water under strong onshore winds.

Wave- and tide-induced currents of sufficient velocity to transport beach detritus are in turbulent flow. Unlike the more or less steady, unidirectional flow of streams, the wave-induced currents take the form of short-period pulsations or oscillations.

Landforms associated with shore currents include all features of shore lines, erosional and depositional. Beaches, bars and spits, wave-cut cliffs, and abrasion platforms all require for their study an understanding of the fundamental principles of pulsating and oscillating water currents. Although the qualitative-descriptive phases of shore processes and forms have been extensively treated by geomorphologists, the quantitative-dynamic aspects have been given little attention by these same workers. In recent years, studies by laboratory personnel of the Beach Erosion Board of the U. S. Army Corps of Engineers and of such hydraulic laboratories as the Fluid Mechanics Laboratory of the University of California at Berkeley, have brought forward much new quantitative information correlating wave characteristics with beach forms and with the attainment of the steady state in the shore profile. (*See* J. W. Johnson, 1949; Krumbein, 1950; Keulegan and Krumbein, 1949; and Willard Bascom, 1951.)

11. Air, the fluid of lowest viscosity among the geomorphic agents, with its content of dust or sand in transport. With the exception of katabatic winds, which are simple downslope flows of cold air formed by inversion near the ground or local upslope winds of thermal origin, winds are not directly related to topographic slopes. They may, however, be traced to gravitational stresses acting upon air masses of differing density, or air pressure. In this respect, the transportational work of wind is unique, because material can be moved upslope or in some direction other than the downslope lines followed by water and mass movements of soil or rock.

Air has extremely low viscosity in comparison with water, hence air engages in turbulent flow at very low velocities. Excepting the thin boundary layer of laminar flow, turbulent flow

may be assumed in air for all practical purposes. Transport of particles of the clay and fine silt sizes takes place by suspension in the turbulent flow; coarse silt and sand move close to the ground in a form of saltation by elastic rebound that is unique among the geomorphic transportational processes.

Landforms and geomorphic processes of turbulent air flow include abrasional and deflational activities, movement of fine-grained particles as dust storms with the eventual production of loess or volcanic ash accumulations, and all dune forms. Geomorphic literature abounds with descriptions of dune forms and with attempts to explain the variations of dune forms with respect to wind strength and direction, vegetative cover, and sand source, but the approach has been descriptive and nonquantitative. To R. A. Bagnold (1941) we owe a great advance in the application of principles of aerodynamics to the study of the movement of sand and the development of drift and dune forms. In a single publication Bagnold has laid the foundations for quantitative and dynamic studies of the erosional, transportational, and depositional work of wind.

Molecular Stress Phenomena

Table 1B outlines the molecular stress phenomena, more conventionally termed the "weathering processes." Molecular stress is principally stress set up by changes of temperature or physical-chemical changes. The important point is that the direction in which the stress acts is independent of gravity and may be distributed in a random manner throughout the rock or soil. Because the source of energy is solar in most of the phenomena listed, the processes are limited to surface or near-surface locations.

1. Direct thermal stresses set up by heating and cooling of rocks. Although all crystalline solids are affected by temperature changes, it is assumed here that unconsolidated granular materials are merely agitated with respect to one another and that actual rupture to form new surfaces of breakage takes place only in hard, crystalline rocks in which mineral grains are tightly knitted by cementation or original bonding of crystal intergrowth. The material is therefore described as an elastic crystalline solid.

Shear stress by temperature changes can be divided into two categories: (a) Nonuniform expansion and contraction by heating and cooling gives rise to shear stresses between adjacent grains of unlike orientation or physical properties. Assuming for the present that the stresses are capable of exceeding the elastic limits of the rock, some granular disintegration, especially of the coarse-grained crystalline rocks, may be attributed to this cause. (b) As a result of cyclic temperature changes at the surface, whether daily, cyclonic, or seasonal, thermal gradients are set up in the rock. Range of temperature at depth for a given heating and cooling cycle at the surface is a negative exponential function of depth. Where surface changes are great and of short period, considerable shear stresses may be set up in the rock, tending to cause rupture in planes approximately parallel with the surface. Exfoliation of hard, fresh rock surfaces is known to occur during intense heating by forest and brush fires and it is possible that other causes of exfoliation may be supplemented by normal solar and atmospheric heating and cooling. Griggs (1936) has applied principles of experimental physics to this problem.

2. Stresses developed during the growth of ice crystals in interstices of rock or of ice lenses and wedges in soil. Open systems are assumed in which atmospheric confining pressures are maintained during crystal growth. Ice crystals of needle-like form, their axes perpendicular to the walls of the opening, exert pressure as water molecules derived from water films are fitted into place at the ends of the crystals. (*See* Taber, 1930; Grawe, 1936.) Bedrock may be shattered by growth of crystals between grains, cleavage fragments, joint blocks, or bedding surfaces. Soil is heaved upward by the growth of ice needles or lenses parallel with the ground surface. The geomorphic effects of rupture of hard rocks by freezing water are especially conspicuous above timber line on mountain summits where talus, felsenmeer, or rock glaciers abound.

3. Stresses exerted by growing salt crystals. This process is essentially the same mechanically as the growth of ice crystals but proceeds

under quite different hydrologic conditions. In arid climates and climates with a long, hot dry season, excessive surface evaporation results in the surfaceward capillary movement of soil and ground water. Upon evaporation, salts, usually carbonates and sulphates, crystallize in the interstices of the soil or rock, exerting pressures upon confining walls. Efflorescence of building stones and brick, granular disintegration of sandstones, and growth of caliche layers and nodules in dry climates are expressions of this process.

4. Shear stresses set up by the adsorption of water by colloids in the rock and soil. Alternating periods of drought and rainfall will cause losses and increases of soil moisture, which in turn cause shrinkage or swelling of clay soils or argillaceous sedimentary rocks. Rupture may occur between grains of the rock or soil, and there is the further possibility of exfoliation of rock shells where moisture penetrates a surface layer, causing it to develop shear stresses along surfaces parallel with the outer surface. Shales and sandstones, as well as igneous and metamorphic rocks, might suffer exfoliation from this cause. In the case of rocks containing feldspars and ferromagnesian minerals, the mineral alteration products, being hydrous clay minerals, would be subject to swelling and contraction, whereas unaltered rock would not. The phenomenon of slaking of shales and bentonites upon exposure is a manifestation of water adsorption.

5. Shear stress set up by contraction of capillary water films between grains of a soil or granular rock. Where discrete capillary films occupy the contact areas of adjacent grains, a shrinkage of the film through evaporation increases the capillary tension of the curved film surface. This in turn causes the grains to be pressed more tightly together and, as a result, the entire mass tends to shrink and to increase in density and strength (Burmister, 1948, p. 80–81). The stresses thus set up may cause shrinkage cracks to form in fine-grained soils in the dry state. In a sandstone, the tension might be sufficient to rupture the rock. Where a mud film on a rock surface dries out, the shrinkage might loosen fragments from the rock surface to which the mud has tightly adhered.

6. Shear stress set up by the growth of plant rootlets in soil or rock. Rupture between grains, joint blocks, cleavage fragments, or bedding layers may result from osmotic pressures of root growth.

7. Expansion of rock upon release of confining pressure, as a result of preexisting elastic strain within the rock. Whether the strain is of diastrophic origin, a relict of mountain-building processes, or a result of solidification of magma under high confining pressures is not fully decided, but the phenomenon of rockburst in quarries, tunnels, and mine shafts is well known and the expansion of slabs upon release from confining rock walls is measurable (Bain, 1931). Large exfoliation domes, as well as widespread development in igneous and metamorphic rocks of exfoliation planes parallel with the hill slopes, have been attributed to expansion accompanying release from confining pressure, or unloading, by erosional removal.

Surficial Creep Phenomena

Not listed in the table but clearly a result of simultaneous action of both gravitational and molecular stresses is the slow surficial creep of soil, weathered rock, or weak bedrock down a slope. Sharpe (1938) discussed the mechanisms of soil creep, but the fundamental law of creep was stated much earlier by Gilbert (1909, p. 345) who wrote: "Whatever disturbs the arrangement of particles, permitting any motion among them . . . promotes flow, because gravity is a factor in the rearrangement and its tendency is down the slope." All types of rupture or intergranular shear of rock or soil treated under the molecular stresses are forms of disturbance of surficial materials in which the directions of stress are either oriented in a random manner throughout the material, or oriented with respect to the rock surface and independent of gravity. Unless there is a slope to the ground surface, no systematic aggregate movement in one direction will occur, but should a slope exist, however faint, a component of gravitational stress will be added to the molecular stresses in the down-hill direction, subtracted in the up-hill direction. Downslope movement is thus cumulative throughout countless minute movements among the grains of the soil. Sur-

ficial soil creep differs from continuous creep, in which flowage deformation progresses because of the plastic nature of the rock. Surficial creep can occur in dry, nonplastic materials with a high degree of internal friction, which would not otherwise flow under gravitational stress alone.

The related phenomenon of downslope creep by rain-drop impact may be noted in connection with surficial creep of soil. Heavy raindrops striking a bare soil surface produce craters from which soil particles are thrown into the air (Ellison, 1950). With perpendicular fall of drops on a horizontal surface, no aggregate movement of soil in any one direction would be expected, but, where a slope exists, the gravitational stress differential comes into play. The trajectories of particles aimed upslope are shorter; those aimed downslope are longer and impart downslope movement to the particles struck. A downslope transfer of particles is thus affected. Unlike the other creep mechanisms, however, gravitational stress on the falling drop, rather than molecular stresses, disturbs the particles and drives this process.

Chemical Processes

Chemical processes (Table 1C) are set apart because they do not directly produce shear stresses, yet are of great importance in landform development. Distinction is made between (1) chemical reaction in which acids ions in soil water or surface water react with mineral surfaces and (2) simple solution or ionization of unusually soluble minerals, such as halite. Removal of the ions in the circulating ground water and in surface runoff constitutes a form of mass reduction that might otherwise be impossible because of resistance of the hard crystalline or glassy minerals to corrosion or to other forms of mechanical reduction.

Chemical reaction or simple solution causes loosening of cementing bonds between grains of a sedimentary rock and selective removal of mineral grains in igneous and metamorphic rocks, thereby weakening the rock and reducing its resistance to both gravitational and molecular shear stresses. Minor forms such as pitting and grooving of rock surfaces are usually described in connection with dissolution of rocks, but the general lowering of land surfaces by removal of mass throughout the soil and weathered rock layer may be of vastly greater significance. In tropical rainy climates, where large quantities of water percolate through the soil and temperatures are most favorable to chemical reaction and solution, the major share of landmass denudation might conceivably be carried out by solution-removal processes. Low slopes or absence of surface slope would not prevent this form of denudation. (*See* discussion of the flow of water downward through permeable materials, Table 1A, No. 7.) The karst landscape features, typified by disruption of normal surface drainage systems, would be assigned to the chemical processes in combination with subsurface percolating or channel flow.

Tectonic and Volcanic Stresses

Although landforms of various types are produced directly by tectonic stresses or by volcanic extrusion, these processes have not been included in this discussion, which takes as its scope the processes of weathering, mass wasting, erosion, transportation, and deposition powered or triggered by solar energy. Whereas these exogene or external processes affect only a thin surficial zone of the landmasses, the tectonic and volcanic processes are deep seated; they involve structural and petrographic changes of enormous masses, along with the transformation of enormous amounts of energy whose source we may regard as internal. Such landforms as fault scarps, initial domes or anticlines, volcanoes, and lava flows are superficial features in terms of the deep-seated changes of state which they accompany. A complete treatment of stresses and behavior of these materials at depth would require inclusion of all of structural geology, petrology, and geophysics into geomorphology. Moreover, initial landforms produced by these deep-seated processes are easily identified and understood as landforms, hence cause no special concern to the student of geomorphology. He considers crustal changes as changes of potential energy of mass with respect to base level, the quantitative values for which are readily calculable by multiplying mass times elevation.

Dynamic Open Systems and the Steady State

Geomorphology will achieve its fullest development only when the forms and processes

are related in terms of dynamic systems and the transformations of mass and energy are considered as functions of time (von Bertalanffy, 1950a; 1950b). Walling himself off in sacrosanct confines of his geological societies and journals, the geomorphologist has paid little attention to the development of thermodynamic principles and their steady infiltration from pure physics and chemistry into sciences of biology, economics, psychology, and political science. True, the Davisian concept of cycle related changes of form to changes of time, but the treatment is not based upon mathematical law, and no thought is given to energy relationships, despite the inviting opportunities for such treatment.

Many of the geomorphic processes operate in clearly defined systems that can be isolated for analysis. A drainage system—whether of water or ice—within the geographical confines of a watershed represents such a dynamic system. A cross-sectional belt of unit width across a shore line or sand dune, or down a given slope from divide to stream channel, would constitute another, more limited, type of dynamic system.

Two major types of thermodynamic systems may be recognized (von Bertalanffy, 1950a): (1) the *closed system* which has a clearly defined boundary through which neither materials nor energy are exchanged, and (2) the *open system* which exchanges either material or energy (or both) with outside environments. The closed system tends to establish an equilibrium in which entropy attains the maximum, available free energy the minimum. An example may be found in the state of water vapor in the air standing above a water surface in a sealed jar. If no heat flows into or out of the jar, the water vapor will attain a certain concentration, maintained without further change of temperature or pressure. The only activity in this system will be exchange of a few molecules between the gas and water. Such a system obviously does not describe a stream or glacier where motion continues with time and material is continually entering and leaving the system.

Form and composition of the open system depend upon the continuous import and export of materials and energy. Normally a time-independent steady state is achieved in which the form remains unchanged but the activity continues. In the case of a segment of a stream in uniform flow, a steady state ensues when the energy developed through the descent of the water is entirely dissipated in overcoming resistance to shear within the fluid and against the channel boundary and to the movement of bed load. The discharge is constant throughout this stretch of stream, and there is no acceleration of motion except what is accountable to changes of channel form, roughness, or slope. In this steady state, the form of the stream is unchanging with time, but should the import of water be cut off the form will be destroyed at once. Open systems require energy from outside to maintain a steady state; equilibrium of the closed system requires none.

Open systems such as streams or glaciers, or cells of living matter, are able to adjust internally to changes in supplies of material or energy from outside. The open system is, in other words, a self-regulating mechanism (von Bertalanffy, 1950a). When a stream is graded, it is in a steady state. If the bed-load supply is reduced or increased, the stream changes its slope so as to readjust to a new steady state. A shore segment whose beach possesses the "profile of equilibrium" is in a steady state with respect to the energy supplied by breaking waves. When the wave characteristics change, the beach profile is altered in slope until a new profile, independent of time, is established.

Mathematical Models in Geomorphology

In attempting to quantify his statements of geomorphic process and form, the geomorphologist has, in general, two types of mathematical procedure open to him, both fruitful. He may, by statistical analysis of experimental and sample field data, derive empirical equations that best state the observed interrelationships between two variables. In its first form, this sort of empirical equation states the degree of correlation between two form elements, both of which may be products of a third and unknown independent variable. For example one might relate drainage density to length of overland flow in a statistical correlation. Although neither quality of terrain is a cause of the other, the degree of correlation is very close because both qualities are controlled by a third, independent factor. Aside from predicting the mag-

nitude of one form element when the other is known, this mathematical statement is of limited value for it does not improve understanding of the genesis of the landform.

In a higher form, the empirical equation may describe a regression in which the independent variable is a force, or time itself, whereas the dependent variable is a landform element. For example, if drainage density were plotted against surface resistivity, a close but inverse relationship would be found. Here resistivity, or resistive force, is a cause; drainage density an effect. Analysis of this type is found in general engineering practice; it is, in fact, the only way in which carefully observed sets of values can be impartially and objectively related.

As a second general procedure, the geomorphologist may formulate, through a type of invention or intuition based upon the sum total of his experience, a relatively simple mathematical model (Rafferty, 1950) which is a quantitative statement of some point of important general theory otherwise definable only in words, qualitatively. The establishment of such mathematical models may be regarded as the highest form of scientific achievement because the models are precise statements of fundamental truths. The two methods—empirical and rational—would tend to converge as time goes on and the fund of information grows. The statistical analyst cannot hope to derive quantitative relationships of general application from small samples because of their inherent variability, but, as his sample data increase and the influences of variables are isolated, his empirical equations tend to approach the status of general laws. New knowledge of the observed influences of variables in turn results in keener deduction on the part of the analyst who is formulating his general mathematical laws by intuitive, deductive mental processes.

To illustrate the formulation of mathematical models in geomorphology, a specific example is offered on the subject of the longitudinal profile of the graded stream. Mr. Samuel Katz of the Lamont Geological Observatory of Columbia University has very kindly worked out the following steps at the suggestion of the author:

We are interested in deriving a time relationship between the elevation of a given point on a graded stream and the horizontal distance of the point from the head of the stream, making only the minimum number of *a priori* assumptions. If y is the elevation, x the horizontal distance measured from the head of the stream, and t the time, we shall take y to be a function of the two independent variables x and t, and shall look for a quantitative expression of this function.

From an analysis of many graded stream profiles (Shulits, 1941; Krumbein, 1937), we feel confident that the relationship between the elevation y and the horizontal distance x of a graded stream is given at a particular time t_0 by an exponential function of the form

$$y_0 = A_0 e^{-k_1 x} \tag{1}$$

(The constant A is determined by the value of y when $x = 0$, and the constant k_1 by the value of y when $x = 1/k$.)

The overall reduction of relief and valley-wall slope steepness will cause a steady reduction of load and a diminishing supply of potential energy in a drainage system. Consequently, the regrading of the master stream becomes increasingly slower. We therefore postulate that, at any given time, the rate at which the stream profile is lowered at a given point is proportional to the slope at that point. Expressed analytically,

$$\left(\frac{\partial y}{\partial t}\right)_x = k_2 \left(\frac{\partial y}{\partial x}\right)_t. \tag{2}$$

From the basic definition of the differential of a function of two variables $y(x, t)$,

$$dy = \left(\frac{\partial y}{\partial x}\right)_t dx + \left(\frac{\partial y}{\partial t}\right)_x dt. \tag{3}$$

Substituting (2) into (3), we eliminate the partial derivative with respect to the time and obtain

$$dy = (dx + k_2\, dt)\left(\frac{\partial y}{\partial x}\right)_t. \tag{4}$$

Equation (4) holds for all values of t, in particular for $t = t_0$. Differentiating (1),

$$\left(\frac{\partial y}{\partial x}\right)_{t_0} = -k_1 y_{t_0}. \tag{5}$$

Substituting (5) in (4),

$$dy = (dx + k_2\,dt)(-k_1 y) \quad \text{or}$$

$$\frac{dy}{y} = -k_1\,dx - k_1 k_2\,dt. \tag{6}$$

Integrating and using the initial conditions that for $x = 0$ (the head of the stream) and

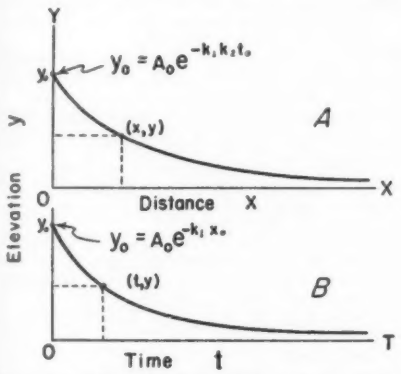

FIGURE 2.—STREAM PROFILE AS A FUNCTION OF DISTANCE AND TIME

A. Profile at a given time, t.
B. Variation in elevation at a point, x, with time.

$t = t_0$, $y = A_0$ (the elevation at the head of the stream), we obtain

$$\ln y - \ln A_0 = -k_1 x - k_1 k_2 t \quad \text{or}$$

$$y = A_0 e^{-k_1(x + k_2 t)}. \tag{7}$$

Equation (7) gives the longitudinal profile of the graded stream both as a function of the time and of the distance from stream head. It has been obtained with no explicit assumption about the variation of slope profile with time.

Figure 2A is a schematic plot of equation (7) for a given value of $t = t_0$, and thus shows the stream profile at a given moment. Figure 2B is a similar plot for a given value of $x = x_0$, and shows the reduction of elevation with time at a point (x_0) on the profile. The exponential relationship of distance to elevation is readily checked by existing profile data, but the rate of reduction of elevation with time will be extremely difficult to verify with field evidence extending back into geologic time. Perhaps controlled model experiments would be required in the latter case.

In summary, the proposed program for future development of geomorphology on a dynamic-quantitative basis requires the following steps: (1) study of geomorphic processes and landforms as various kinds of responses to gravitational and molecular shear stresses acting upon materials behaving characteristically as elastic or plastic solids, or viscous fluids; (2) quantitative determinations of landform characteristics and causative factors; (3) formulation of empirical equations by methods of mathematical statistics, (4) building of the concept of open dynamic systems and steady states for all phases of geomorphic processes, and finally (5) the deduction of general mathematical models to serve as quantitative natural laws. The program is vast and qualified investigators few, but we are already a half-century behind if development is to be measured against chemistry, physics, and the biological sciences. The need for rapid dynamic-quantitative advances is, therefore, all the more pressing.

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Department of Geology, Columbia University, New York, N. Y.

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PROBLEM OF GLACIATION IN THE WESTERN ISLANDS OF ARCTIC CANADA

By John L. Jenness

Abstract

This paper deals with the Western Islands of the Canadian Arctic Archipelago (that part of the Archipelago which lies west of approximately the 95° meridian of West Longitude). It directs attention to some important aspects of the study of former glaciation in this region, and attempts to shed light on the extent and character of this glaciation.

The following major conclusions have been reached: The islands south of Viscount Melville Sound (with the possible exception of Somerset Island) clearly appear to have been glaciated by ice moving off the continental mainland. Evidence suggests that this ice advanced into the Archipelago more than once but there is no indication that it ever extended north of Viscount Melville Sound. In the northern islands the highest upland surfaces may have been occupied by local glaciers. Ice flowage is suggested at Resolute Bay on Cornwallis Island, in the Deer Bay region on Ellef Ringnes Island, and may be the cause of furrows and striae on bedrock near the head of Liddon Gulf, Melville Island. The negative character of the photographic evidence in the other northern islands does not preclude the possibility of former local glaciation. Dormant or semidormant snowfields or ice caps similar to the one now on Meighen Island may have been widely distributed at one time.

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Introduction

Prior to the summer of 1950, when the Royal Canadian Air Force completed its aerial survey of the islands in the Canadian Arctic, no single traveller could acquire first-hand knowledge of more than a very small part of them. Whoever undertook to study the geology, topography, or other characteristics of the region as a whole was compelled to base his conclusions largely

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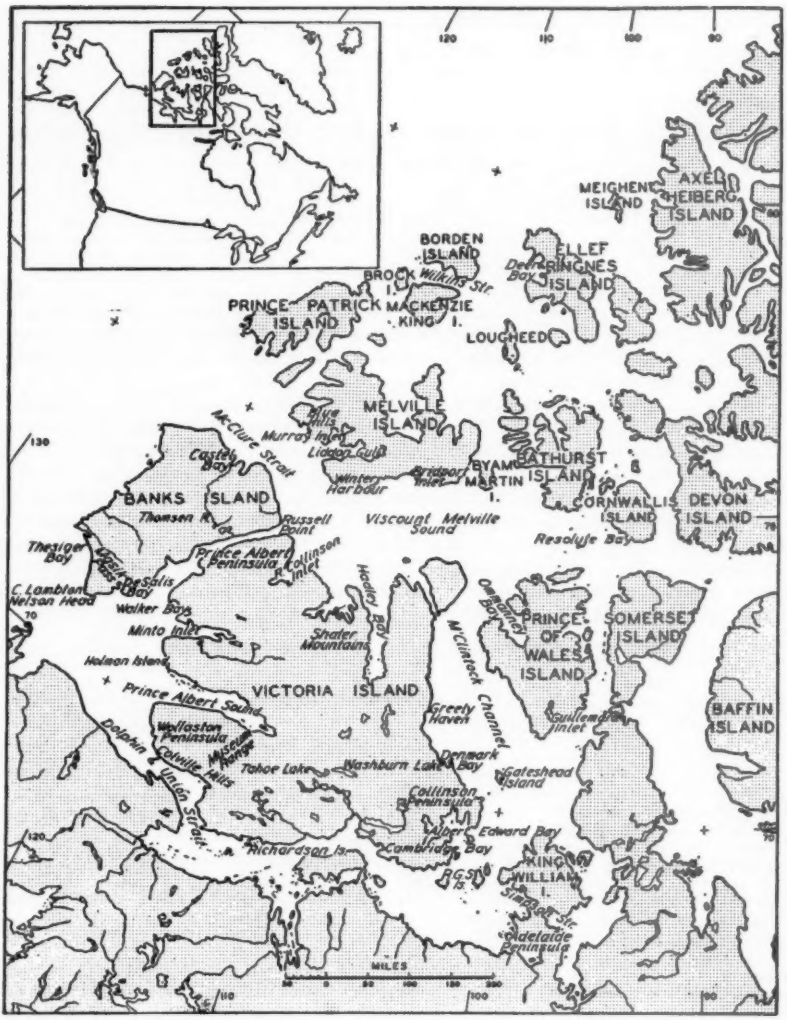

FIGURE 1. LOCATION OF THE WESTERN ISLANDS OF THE CANADIAN ARCTIC ARCHIPELAGO

on unconfirmable data gathered from scattered localities by other observers. It is still physically impossible for one man to visit all of the many islands, but the opportunity to make valid generalizations has immensely improved. All of the Arctic can now be seen and photographed from the air; and observers can both land at many different points on the ground, and check their own and earlier observations by studying the terrain on aerial photographs.

EARLIER VIEWS REGARDING GLACIATION

The widely divergent views that have been advanced on the subject of former glaciation within the western islands of the Arctic Archipelago reflect the almost complete lack heretofore of detailed information. These views run the whole gamut from complete glaciation to complete lack of glaciation; and authors favoring the former have disagreed among them-

selves as to whether the glaciation originated within the Archipelago itself or entered it from the continental mainland. The most recent writer on the subject, A. L. Washburn, (1947) combined careful field observations on Victoria Island and adjacent regions with an exhaustive survey of the earlier literature. He proved that at least a part of the Archipelago had been glaciated during the Pleistocene, and suggested that glaciation may have been widespread over the whole of the region.

Washburn's main conclusions were the following:

1. The mainland coast and at least most of Victoria Island were glaciated.
2. Till-like deposits containing striated stones and erratics (some of which appear to have originated *in situ*) are present in southern Banks Island. In view of this, and of the fact that the coast of Victoria Island to the east was unquestionably glaciated, the southern portion of Banks Island also is assumed to have been glaciated.
3. Crescentic fractures, crescentic gouges and well-developed striae strongly suggest glacial movement from the southeast into the Royal Geographical Society Islands.
4. Because Adelaide Peninsula on the continental mainland opposite King William Island was demonstrably glaciated by ice moving northward or northwestward, King William Island itself must have been glaciated.
5. The widespread occurrence of erratics in the Archipelago may indicate a widespread glaciation, although the value of erratics in determining glaciation is questionable in regions where ice rafting may have operated.
6. The widespread emergence of the Arctic Archipelago, demonstrably post-glacial in many places, seems best explained by isostatic adjustment to removal of ice load. This suggests glaciation of the whole, or almost the whole of the Western Arctic. (Washburn cautions however, that while emergence suggests glaciation, our inadequate knowledge concerning the areal extent of isostatic adjustment and its influence on adjacent regions, together with the possibility of diastrophic movements unrelated to glaciation, make its evidence indeterminate.)

At this stage the matter rested in 1947. Clearly, the extent and character of former glaciation cannot be finally settled without many more ground observations. We may facilitate its ultimate solution, however, if we review the various lines of physiographic evidence that have been used to identify glaciated land in Arctic Canada, established which of these lines is valid and to what extent, and, using material that has become available since Washburn wrote,[1] outline our most recent information concerning the probable area and character of glaciation in this region.

Physiographic Criteria of Glaciation

Changes in Relative Level of Land and Sea

Aerial photographs reveal deeply indented, highly irregular shore lines on all islands in the Canadian Western Arctic that have elevations of 1000 feet or more, and in many places one can recognize the drowned coasts indicative of submergence. Yet even more widely spread are the evidences of recent emergence, visible on large and small islands alike, thereby indicating that they have risen as a unit. Banks, Somerset, Cornwallis, Bathurst, Melville, Prince Patrick, Borden, Mackenzie King, Lougheed, and Ellef Ringnes islands all carry deeply-incised streams that arise on interior uplands and possess canyons with the V-shaped profiles that distinguish stream erosion from glacial erosion. The freshness of these streams implies a considerable recent change in land level relative to the sea. So too do the streams on these islands that originate at a lower level, and the streams on islands of more limited height such as King William and Byam Martin; for they have all failed as yet to cut deep channels, but give the impression of flowing over an initial surface. Emergence exceeding 500 feet has actually been demonstrated at Cambridge Bay, Victoria Island, and at Reso-

[1] Viz. The aerial photographs of the Archipelago taken by the Royal Canadian Air Force, and my own observations on Banks, Melville, and other islands during the summers of 1948 and 1949. The present paper deals only with the glaciation of the western part of the Archipelago—that part lying west of approximately 95°W., as depicted in Fig. 1.

lute Bay, Cornwallis Island.[2] All parts of the Archipelago may not have risen equally, yet for the region as a whole a rise of 400 to 500 feet and perhaps more would not appear excessive.

Now an uplift of this magnitude is highly significant. Throughout the southern part of the archipelago, eskers and other depositional material provide abundant evidence of glaciation; but north of Viscount Melville Sound no ground moraines or glacial outwash have been recognized. Those who advocate the glaciation of that area rely largely on the presence of foreign erratics, but foreign erratics within several hundred feet of present sea level (i.e. within the zone of recent emergence) may well be the products of ice rafting, not of glaciation. Even if the seas in the southern part of the Archipelago were rather more ice-free during the period of emergence than they are at present, as Washburn suggests, nevertheless it seems very unlikely that any post-Pleistocene amelioration of climate, and of sea-ice conditions, can have been so great as to spare the shores of the northern islands from ice action, whatever may have been the case in the south. In this connection, it is well to remember that movement of erratic material by sea ice may have been greater in the past than it is today, especially if the climate was somewhat more moderate, for the larger expanses of sea between the islands may have given ice a greater freedom of movement than it enjoys now, and permitted a wider transportation of ice-rafted material. For these reasons, erratics alone should not be accepted as reliable evidence of glaciation except at elevations above the former marine limit. Within the lower several hundred feet of present sea level, they can be considered valid only where they are associated with other more positive indicators of glaciation. I know of no foreign erratics north of Viscount Melville Sound that are above the marine limit, or of more positive indicators of ice-sheet glaciation either above or below it.

Like foreign erratics, striated surfaces also may not be conclusive evidence of glacial erosion. Even in West Virginia, which has never been glaciated, stream ice is capable of forming grooves "running as straight as glacial grooves for a length of 40 or 50 feet" in the underlying bed-rock (Washburn, 1947, p. 46, 48). In the Arctic too, I have seen boulders marked with well-developed striae that were probably caused by pack ice and not glaciation; for pack ice can exert amazing pressure and erosive power, as Stefansson's (1922, p. 509) description of its action at Point Barrow, Alaska, and elsewhere vividly illustrates:

"Sometimes the ice heaps up on the land and thousands of tons of it are shoved hundreds of feet inland and twenty or thirty feet above the level of high tide." (Stefansson, 1922, p. 509).

Striae should therefore be accepted as positive evidence of glaciation only when (1) they are well-developed at elevations above the maximum marine limit or (2), at lower levels, when they are parallel through long distances over a large surface and are associated with well-smoothed bedrock and other features that also suggest glacial erosion.

McMillan (1910, p. 411, 461) is the only author who mentions striae north of Viscount Melville Sound. He noted markings on the surface of hard white sandstone on the southern side of Liddon Gulf, Melville Island, about 2 miles from its head. This locality is low-lying and well below the marine limit. He found no glacial markings elsewhere, and makes no mention of other glacial evidence, despite his prolonged stay on Melville Island. It seems unreasonable, therefore, that the presence of this one scratched surface, unaccompanied by other evidence, should be accepted as proof that the island was once subjected to continental glaciation. McMillan himself, indeed, thought that the markings might have been caused by local glaciers, a possibility that cannot be dismissed without further investigation, as there is now independent evidence suggesting local glaciation at one or two other places in the northern islands.

Depositional Landforms

Eskers offer the most valid evidence of glaciation in the Western Arctic, and have the advantage of being clearly recognizable, as a rule, both from the air and on aerial photo-

[2] See discussion on Mt. Pelly, Cambridge Bay, Victoria Island, in Washburn (1947); data on Resolute Bay is from R. L. Nichols, Personal communication.

graphs (Pl. 1, fig. 1). They are positive evidence of glaciation even when they lie far below the marine limit. If found above that limit, they may represent the moulds of tunnels; but in the western Arctic Islands I have observed no eskers except within several hundred feet of present sea level.

Victoria Island contains many eskers, the majority on the southern lowland section that extends from Albert Edward Bay on the east coast to Prince Albert Sound in the west. These are all well preserved, and associated with terrain containing a superabundance of lakes. Photographs reveal other forms resembling eskers, but in a much poorer state of preservation, in the central part of Wollaston Peninsula north of the Colville Hills, and in Prince Albert Peninsula on the northern side of the belt of trap-sedimentary rocks that extends from Minto Inlet to the Shaler Mountains and Hadley Bay. While lakes are numerous in both these localities, they are far less common than in the southern lowland. King William Island contains many eskers with well defined shapes (Pl. 1, fig. 2). They can be recognized in almost any part of the island, and are associated with a terrain containing an abundance of lakes comparable to the lowland part of Victoria Island. A few forms resembling eskers are visible on aerial photographs of Prince of Wales Island, but only in the lake-filled, lowland southern half; they are rare compared with the numbers found on Victoria and King William Islands. The majority are situated in the Ommanney Bay region, and there, as elsewhere on the island, lie close to present sea level. I recognized no eskers on Banks or Somerset Islands, or on any of the islands north of Viscount Melville Sound.

Land Patterns

On many islands, swales lying between a succession of former beach ridges along a particularly gentle stretch of coast will accentuate the contour patterns created by these beaches. I have seen such swale, beach-ridge patterns on both sides of Dolphin and Union Strait, for example, and they are apparently quite conspicuous, also, in the vicinity of Guillemard Inlet, Prince of Wales Island. But a quite different pattern prevails on the three esker-rich islands, Victoria, King William, and Prince of Wales, and on them alone—a pattern apparently caused by glaciation rather than by uplift of the land or some peculiarity in the character of the underlying bedrock. In some places, as in west-central King William Island, and along and inland from the eastern coast of Victoria Island, the topography is given a definite trend over long distances by parallel arrangements of elongated lakes and elongated ridges. Fortier (1948, p. 7, 8) describes it thus:

> "An interesting feature . . . is a belt of northwesterly oriented drumlinoids extending from the mainland south of Simpson Strait and covering the western part of King William Island, Collinson Peninsula on Victoria Island and northward of Denmark Bay to Greely Haven. Gateshead Island is also part of this belt. The drumlinoids are elongated ridges of drift, longer than common drumlins, although locally the forms may be called drumlins. The intervening valleys and lakes are elongated parallel with the drumlinoids, as are parts of shores of larger lakes and the peninsulas and islands, and islands off the coast."

On aerial photographs, this linear topography can be traced beyond the limits given by Fortier. It is found on the continental mainland, where it crosses from Adelaide Peninsula through a series of elongated islands in Simpson Strait into King William Island; it extends north of Greely Haven, on the eastern side of Victoria Island, as far as the end of the peninsula between M'Clintock Channel and Hadley Bay, and may be present as well on the unnamed island lying just beyond that peninsula; and it crosses southwestern Prince of Wales Island. The pattern runs from southeast to northwest, but, being arcuate rather than straight, its direction is more nearly south-southeast by north-northwest as it crosses from Adelaide Peninsula into King William Island, and east-southeast by west-northwest where it leaves the western side of Prince of Wales Island and continues on along the east coast of Victoria Island.

A linear topography characterized by elongated lakes gouged out of a flat landscape, but without drumlinoids, prevails across the whole or nearly the whole lowland of Victoria Island (Pl. 2, fig. 1). In some places it is remarkably distinct; in others it tends to disappear, as a result, perhaps, of inadequate drainage. The same pattern of elongated lakes occurs also, though not so clearly, in those parts of King

William Island that lack the drumlinoids. Since it is identical in character and direction with the drumlinoid-lake pattern, the two presumably owe their origin to the same glacial movement—a movement that probably produced also the crescentic fractures, crescentic gouges, and long and regular NW-SE. striae that Washburn noted in the Royal Geographical Society Islands (Washburn, 1947, p. 56).

On Victoria Island a line of lakes extending inland from the south coast creates a second trend, elongated roughly north to south. The two trends overlap in the center of the island, close to Tahoe Lake; and the shapes of the lakes in this region reflect both orientations (Pl. 2, fig. 2). They strongly suggest two stages of ice movement on Victoria Island, and may indicate separate sources of accumulation on the mainland. No trend of any kind is distinguishable in northwestern Victoria Island beyond the belt of trap-sedimentary rocks that extend from Minto Inlet to Hadley Bay; or in Somerset Island apart from a small string of lakes, oriented north and south, along the contact between a granitic upland in the west and a sedimentary lowland in the center. None exists in Banks Island, or in any island north of Viscount Melville Sound. I believe that the directions of these linear patterns in Prince of Wales, King William, and Victoria islands, combined with the eskers and other evidences of glaciation in the same islands, prove that this part of the Archipelago was glaciated by ice moving into it from the continental mainland. Figure 2 shows what I conceive to have been the maximum extent of continental glaciation in these Western Arctic Islands.

Drainage Patterns

Hobbs (1945, p. 549–551) has used a certain aspect of the drainage patterns in the islands to determine the limits of the advance of Wisconsin ice into the Archipelago. According to him, "the area covered by the latest glaciation is everywhere characterized by the occurrence of a peculiar type of glacial lake with many bays and islands, and these lakes are found in great numbers—in myriads wherever the relief is low and the glacial topography dominates the landscape." Such lakes, which have a widespread distribution over the northern mainland west of Hudson Bay, are numerous also over the whole of King William Island, throughout nearly all Victoria Island, in southwestern Prince of Wales Island, and in eastern and southeastern Banks Island (Pls. 1, 2). The lakes strung out on Somerset Island along the contact between the granite upland and the central lowland may belong to this group also,

PLATE 1.—CONSPICUOUS ESKERS IN THE SOUTHERN PART OF THE ARCHIPELAGO

FIGURE 1. LOOKING WEST ACROSS THE CENTRAL, LOWLAND SECTOR OF VICTORIA ISLAND

The branching esker is proof of former glaciation, and indicates a retreat of the ice towards the east (towards the foreground of the photo). Note the multitude of lakes and ponds. (Photograph, courtesy of R.C.A.F.)

FIGURE 2. LOOKING WEST ACROSS KING WILLIAM ISLAND FROM APPROXIMATELY ITS GEOGRAPHICAL CENTER

Note esker across the foreground; also the multitude of lakes and ponds, and the indistinct trend which crosses the photo from lower left to upper right (in a southeast-northwest direction). (Photograph, courtesy of R.C.A.F.)

PLATE 2.—LINEAR TOPOGRAPHY CHARACTERIZED BY ELONGATED LAKES

FIGURE 1. LOOKING WEST ALONG THE CORONATION GULF COAST OF VICTORIA ISLAND

A distinct topographical trend which crosses the photo from lower left to upper right seems caused by glaciation and probably represents the direction of glacial movement. The light-colored markings are structural terraces. (Photograph, courtesy of R.C.A.F.)

FIGURE 2. LOOKING WEST ACROSS THE NORTHEASTERN PENINSULAR SECTION OF VICTORIA ISLAND WITH THE HEAD OF HADLEY BAY VISIBLE IN THE BACKGROUND

The deeper lakes (ice covered) are oriented north-south, and represent a directional trend more clearly visible farther south. The multitude of shallow ponds and lakes show an indistinct trend from east-southeast to west-northwest, a trend which is very distinct along much of the eastern coast of the island. Both trends are thought to be caused by glaciation and seem to indicate ice movement across this part of the island on more than one occasion. (Photograph, courtesy of R.C.A.F.)

FIGURE 1

FIGURE 2

CONSPICUOUS ESKERS IN THE SOUTHERN PART OF THE ARCHIPELAGO

FIGURE 1

FIGURE 2

LINEAR TOPOGRAPHY CHARACTERIZED BY ELONGATED LAKES

FIGURE 1

FIGURE 2

TERRAIN FEATURES NORTH OF VISCOUNT MELVILLE SOUND

FIGURE 1

FIGURE 2

FIGURE 3

CONSPICUOUS TERRAIN FEATURES IN THE GLACIATED SOUTHERN PART OF THE ARCHIPELAGO

although their distribution shows strong accordance with an abrupt change in structure.

The transition from country abounding in lakes to more northerly country where lakes are almost totally absent tends to be abrupt. Northern Banks Island, the whole northern and eastern parts of Somerset Island, and much of the northern half of Prince of Wales Island are virtually free of lakes, while north of Viscount Melville Sound lakes are absent altogether (Pl. 3, fig. 1).[3]

There is another anomaly in the drainage patterns of the Western Arctic that seems related to former glaciation. North of Viscount Melville Sound all the major streams are of meltwater origin, and none possess the characteristics of maturity, or show any indication of having ever been diverted from their initial courses, even by uplift.[4] Aerial photographs suggest a widespread superposition of drainage (Pl. 3, fig. 2), and also multi-cycle drainage in some of the valleys. On the other hand, in those southern parts of the Archipelago that bear definite marks of glaciation, the drainage is very disorganized, just as it is also over a large part of mainland Canada. Many of the myriad lakes have no visible outlets, but overflow from one to another in so indefinite a manner that it is difficult to detect either the direction of drainage or its ultimate destination. Victoria Island has very few major rivers, considering its size, and even they are shallow, showing that they tap almost none of the many lakes which hem them in along their courses. All King William Island, and the southern part of Prince of Wales Island, have extremely youthful drainage similar to that of Victoria Island. Such drainage is a characteristic product of glaciation, and glaciation is amply attested in these islands by eskers and other evidence.

Nevertheless, two regions in the southern part of the Archipelago are exceptional. In the northern and northeastern sections of Somerset Island, drainage patterns resemble those of the

[3] No water bodies of any kind could be distinguished on tri-camera mapping photographs taken from an altitude of 20,000 feet, except some small ponds concentrated in three or four localities, viz. around Resolute Bay on Cornwallis Island, on Melville Island at the heads of Bridport and Murray Inlets, and on Brock Island near its western shore.

[4] Thus many of the streams which flow off Borden Island into Wilkins Strait possess deeply incised meandering courses which show that they have not been diverted by uplift of the coastal plain on which they originated.

PLATE 3.—TERRAIN FEATURES NORTH OF VISCOUNT MELVILLE SOUND

FIGURE 1. LOOKING EAST ACROSS BORDEN ISLAND

Lakes are absent from this island and from all others north of Viscount Melville Sound. Hobbs has used this absence of lakes as his criterion for believing that the northern part of the Archipelago escaped glaciation altogether. The dendritic character of the drainage gives the impression of streams flowing upon an initial surface, and suggests that the island may have been beneath sea level until recently. Only in the south-central part of the island (not distinguishable in the photograph) does the drainage have a different character; here tortuously meandering streams have been incised to a considerable depth, suggesting that they once flowed on a coastal plain and have retained their initial courses despite uplift. (Photograph, courtesy of R.C.A.F.)

FIGURE 2. LOOKING EAST ACROSS NORTHERN BATHURST ISLAND, EAST OF MAY INLET

Like other islands north of Viscount Melville Sound, Bathurst contains none of the lakes which have such a widespread occurrence in the southern part of the Archipelago. Note piracy, or more probably superposition of drainage across folded structures, by tributaries of the master stream which drains towards foreground of photograph. (Photograph, courtesy of R.C.A.F.)

PLATE 4.—CONSPICUOUS TERRAIN FEATURES IN THE GLACIATED SOUTHERN PART OF THE ARCHIPELAGO

FIGURE 1. IRREGULAR, ROUGH TOPOGRAPHY, SEEMINGLY COMPOSED OF UNCONSOLIDATED MATERIALS, IN VICINITY OF MT. BUMPUS, VICTORIA ISLAND

(Photograph courtesy of A. L. Washburn.)

FIGURE 2. MOUNT PELLY, VICTORIA ISLAND, SHOWING EMERGED STRAND LINES EXTENDING ALMOST TO ITS SUMMIT

(Photograph courtesy of A. L. Washburn.)

FIGURE 3. GLACIALLY SMOOTHED AND STRIATED DIABASE NEAR HOLMAN ISLAND (POST) VICTORIA ISLAND

(Photograph courtesy of A. L. Washburn.)

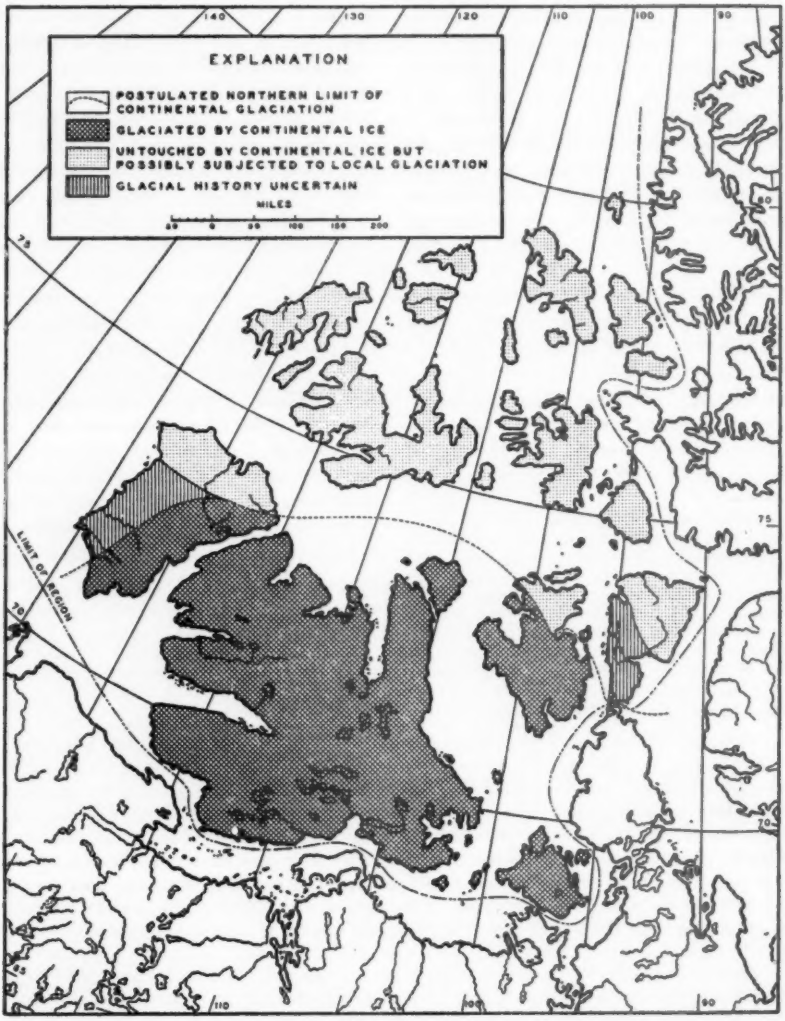

FIGURE 2.—EXTENT AND CHARACTER OF FORMER GLACIATION

northern part of the Archipelago more than those of the southern (Jenness, 1952, Fig. 2). On northern Banks Island, again, stream canyons and V-shaped valleys indicate fluvial action rather than glacial erosion. On Banks Island, too, we find one of the longest rivers in the entire Archipelago, perhaps the longest in its western part, the north-flowing Thomsen River which outlets into Castel Bay after draining half the island. This river meanders on a flood plain, a condition of maturity that is quite exceptional in the region, where uplift has opened a new cycle of erosion. Despite its great length, the river looks quite shallow from the air and in photographs; it meanders in and out among the myriad of lakes and swamps that cover the broad flood plain in the upper half of its course, but seemingly taps very few

of them. Its position, direction of flow, and misfit character all suggest that it may be the descendant of an ice-margin stream dating back to a time when ice occupied Victoria Island to the eastward. Masik Pass, which crosses from DeSalis Bay to Thesiger Bay in the southern part of the same island, also contains underfit drainage. I suspect that this pass marks the limit of an ice lobe that invaded the island from the south.

Other Features Indicative of Glacial Movement

Unclassified depositional forms of various kinds have a widespread distribution in Victoria and King William Islands, and are found also in eastern Banks Island and the southwestern part of Prince of Wales Island. At many places in this part of the Archipelago, I noted isolated hills and belts of hilly and hummocky terrain, all seemingly composed of unconsolidated materials (Pl. 4, fig. 1). The same features are easily distinguishable by their color on tricamera aerial photographs. All seem to be the product of former glaciation, because foreign erratics have been collected on some of their summits high above the marine limit (e.g., on top of the 1700 foot high Mt. Bumpus on Victoria Island. *See* Washburn, 1947, p. 52–53).

These various depositional forms show superficial similarities to various kinds of glacial deposits recognized in glaciated country elsewhere. Dr. R. J. Lougee has pointed out to me that photographs taken by Washburn (1947, Pls. 11 and 17) of Mt. Pelly, an isolated flat-topped hill 675 feet in elevation near Cambridge Bay, Victoria Island, show that this huge deposit of gravels and sands closely resembles such pro-glacial deltas in New England as the "Promenades" at Portland, Maine (Pl. 4 fig. 2). The Colville Hills and Museum Range of Wollaston Peninsula, Victoria Island, resemble terminal or medial moraines, while there are kame-like deposits in both these ranges, on both northern and southern sides of Prince Albert Sound and in eastern and southeastern Banks Island. Ground moraine without distinctive topographic form is widespread in Victoria and King William islands and seems to be present also on Prince of Wales Island. Washburn (1947, p. 49) describes much of the surface at De Salis Bay, Banks Island, as being mantled by a till-like deposit containing striated stones and erratics, and I have seen similar material at a number of landing-points along the eastern side of the same island between De Salis Bay and Russell Point. At two places on Victoria Island are large sand plains that may represent glacial outwash plains. One is in the center of the island about 30 miles north of Washburn Lake; the second, much more minutely dissected, at the head of Richard Collinson Inlet.

The surfaces of these forms have been more or less modified by marine processes and solifluction, so that careful and detailed study will be necessary before they can be exactly identified. Whatever their precise character, they are almost certainly products of glaciation and no features resembling them in either color or appearance show up on tricamera aerial photographs of any island north of Viscount Melville Sound, or of the uplands of northern Banks Island, northern Prince of Wales Island, or northeastern Somerset Island; nor did I discern any, either from the air or on the ground, during my several flights in those areas. Their apparent absence, of course, does not definitely prove that these northern regions remained untouched by continental ice. It strongly suggests their immunity, however, especially as typical glacially deposited forms are generally prominent at the maximum point of advance of a continental ice-sheet, and immediately beyond it.

Landforms associated with continental-type glaciation are most clearly recognizable from the air and on aerial photographs when they are located in a region of slight relief and uniform topography; in more rugged terrain, the search for evidence of ice-sheet glaciation has to be carried out largely on the ground. In the Western Arctic Islands, nearly all the glacially produced features discussed thus far are most prominent on lowlands, where horizontally lying bedrock creates a broad regional uniformity of surface. The more rugged districts, such as the uplands of trap-sedimentary sequence rocks on Banks and Victoria islands and the granite upland of Somerset Island, are still virtually unknown. However, studies conducted recently at two or three localities in these areas sub-

stantiate the idea of widespread glaciation over the southern part of the Archipelago. Thus Washburn has indicated that continental ice overran the trap-sedimentary sequence terrain at the Richardson Islands on the south coast of Victoria Island, and similar terrain in the vicinity of Holman Island, Minto Inlet (Pl. 4, fig. 3), and Walker Bay on its west coast. In these localities, stoss-and-lee topography is common, and glacial smoothing and striae are seen frequently, especially on trap rock (Washburn, 1947, p. 50). The same features characterize the trap rock terrain in the vicinity of Washburn Lake, where they exhibit divergent directions of movement.

Aerial and ground observations in 1949 revealed similar positive evidence of former glaciation in the trap-sedimentary upland of southernmost Banks Island. About 40 miles north of Cape Lambton, at the northwestern extremity of the belt of trap-sedimentary rocks which produces cliffs as far around as Nelson Head, we found distinct striae, grooves, and polished surfaces in large exposures of trap that were characterized by stoss-and-lee slopes. These slopes and markings indicated a glacial movement from the south or south-southwest. Although the highest exposures were only about 250 feet above present sea level, and therefore below the marine limit, the evidence is not on that account invalidated because (a) the striae, grooves, etc. retained a consistent direction over large surfaces, this direction itself being contrary to that of any probable ice-shove, and (b) it was supported to a height of 1200 feet (that of the highest hill in the region) by finds of erratics which were foreign to Banks Island.

Within the same trap-sedimentary upland, we made another landing on a small rock-basin lake just inland from Nelson Head. The lake was one of several in the immediate vicinity, all at about the 2000 foot level and therefore high above the marine limit. Along with an abundance of angular, frost-weathered rock debris of presumably local origin, the slopes surrounding the lake contained numerous erratic stones and boulders, of which certain ones are known *in situ* only on the continental mainland. Now the character of the topography along the coast between Nelson Head and Cape Lambton suggests that the ice which carried these erratics reached Banks Island from the south, a direction which accords well with the movement suggested by the striae and other evidence northwest of Cape Lambton. We may conclude, therefore, that the southern coast of Banks Island was invaded by continental ice that arrived directly from the mainland without first crossing Victoria Island.

Local Glaciation

Glaciers of both ice-cap and mountain type exist at the present time in Baffin, Devon, Ellesmere and Axel Heiberg islands in the eastern part of the Archipelago; but despite the presence of a tiny snowfield capping Meighen Island, and possibly others surmounting the Blue Hills of Melville Island, no active glaciation is visible today in the Western Arctic. Yet in a region such as this, a decrease in summer temperature without change in precipitation, or a considerable increase in precipitation without accompanying temperature increases, would permit the development of permanent or semi-permanent snowfields of the Meighen Island type on the domed central part of Cornwallis Island and on other upland surfaces where there are now none; it might even cause sufficient accumulation to turn the snowfields into glaciers. It is well known that the climate was more severe during the Ice Age than at present, and that it has undergone several fluctuations since that time. This in itself would suggest the possibility of considerable local glaciation at one or more periods in the past, a possibility strengthened by the post-Pleistocene isostatic rise of the entire Archipelago, even though only its southern portion may have been overrun by continental ice.

Have we then any positive evidence for the one-time existence of glaciers in the Western Arctic? I have found no indications at all in the literature, but several hints of it on the aerial photographs. In the islands south of Viscount Melville Sound which have been largely overrun by continental ice, the photographs suggest former local glaciation at one and possibly two places: in the Shaler Mountains of northern Victoria Island, and perhaps also on the upland above Nelson Head in southern Banks Island. In the Shaler Mountains, rock-basin lakes, cirques, and U-shaped

valleys all seem to be not only present, but well-developed. These mountains, however, are Precambrian trap-sedimentary formations containing visibly faulted and tilted structures, so that what appear on the photographs to be ice-erosive features may after all be structural rather than erosional.

In this area local glaciation can be readily confirmed or disproved by ground observations. Elsewhere evidence may be difficult to find; yet even its apparent absence would not exclude all possibility of local glaciation. Dormant glaciers such as that now present on Meighen Island lack the eroding power of active glaciers, and leave little or no traces when they disappear.[5] They contribute only indirectly to the erosion of the landscape by feeding meltwaters into the streams that emerge from them and by saturating the bare ground around their margins, thus promoting mass wastage.

The photographs show no evidence of continental glaciation in the islands north of Viscount Melville Sound and an almost equal dearth of evidence suggesting local glaciation. At only one place did I notice some prominent features that might have been caused by the latter—on the long peninsula which juts out between Deer Bay, on Ellef Ringnes Island, and the large unnamed bay lying immediately north of it. There, close together, a number of deep amphitheatre basins resembling cirques open on to the present coast. Nowhere else in the northern islands did I identify any prominent features which could not be explained without recourse to glaciation, whether local or continental. Streams which emerge from the highest uplands—as from the Blue Hills of Melville Island—exhibit the V-shaped profiles indicative of stream erosion, but they throw no light on the glaciated or unglaciated nature of the land surfaces which they drain. The plateau-like upland surfaces prevailing in western Melville Island, on the southeast side of Prince Patrick Island, and in Ellef Ringnes Island (apart from the peninsula just mentioned), reveal no rock-basin lakes, cirques, hanging valleys, or other features suggestive of glaciation. Even those bays whose fjordlike appearance caught the attention of early explorers seem attributable to the drowning of the lower valleys of multicycle streams rather than to glaciation. Everywhere the photographic evidence conveys the impression of an intimate relationship between surface and structure, and discounts any widespread distribution of glacial topography.

Nevertheless, several of the northern islands share a common characteristic that may have an important bearing on the question of local glaciation. On Cornwallis, Bathurst, Melville and Ellef Ringnes islands aerial photographs suggest the presence of superimposed drainage. They show, also, on certain of the least denuded uplands of these islands (e.g. along certain watersheds) some pale-gray deposits which I believe to be remnants of a former blanketing material that has been largely or entirely removed by erosion from other localities. Specially noticeable is the fact that this gray mantle persists on the domed central part of Cornwallis Island, whose eastern and southern coastal sections have suffered heavily from denudation. Now at Resolute Bay, on the southeast coast of that island, Nichols found eight separate outcroppings of planed, smoothed and striated bedrock that appear to indicate a flowage of glacier ice from the center of the island toward its margins. The outcroppings lie "at the strand line where the weathering processes have not had time to destroy them, as they have been underwater most of the time since formed" (Nichols, R. L., personal communication); and the striae, some of which extend for a distance of 100 feet, point from the center of the island in an approximately north-south direction along the strike of the rocks. In this place, then, we seem to have definite evidence of local glaciation at some undetermined time in the past.

Conclusions

The foregoing discussion leads to these tentative conclusions concerning the extent and character of former glaciation in the Western Islands of the Arctic Archipelago:

Glacial erosional evidence, the presence of

[5] Mathes (1942, p. 154–5) has estimated that in temperate latitudes glacial ice must be at least 100 to 150 feet thick before it will commence to flow, and concluded that in polar regions it would require much greater pressures, and therefore greater thicknesses.

foreign erratics above the marine limit, and characteristic glacially moulded topography indicate that the southernmost part of Banks Island was invaded by ice moving northwards from the continent. A line of hills which seem composed of unconsolidated materials may mark the terminal moraines of this ice, and Masik Pass the channel of a former ice-marginal stream. Thomsen River would appear to be the descendant of a second ice-contact stream and mark the western limit of an ice lobe which reached Banks Island from Victoria Island.

Victoria Island was once entirely or almost entirely overrun by ice from the continental mainland. This is indicated by a widespread distribution of eskers, ground moraines, and other features, and also by linear patterns, attributable to ice movement, which stretch across much of the island, but particularly along its entire eastern coast and part of the south coast. The overlapping of linear patterns with different orientations, and the presence of both well-preserved and badly weathered eskers, the former occurring within a part of the island that contains a multitude of lakes and the latter only where lakes are much less numerous, both show that ice moved in several directions on Victoria Island, perhaps on more than one occasion and from more than one place on the mainland. At Washburn Lake, erosional features offer clear evidence of glacial movement from two overlapping directions.

The badly weathered eskers located in such isolated parts of Victoria Island as the interiors of Wollaston and Prince Albert peninsulas, and the comparative scarcity of lakes in those two regions and in northernmost Victoria Island, suggest that the island underwent two glaciations, an earlier one that covered the whole island, and a later one that left the well-preserved depositional forms and the myriads of lakes. Judging by the drumlinoids, the ice of the later movement crossed the island in a west-northwest direction from the east coast, whereas the earlier movement was from south to north. The Museum Range may represent a terminal moraine of the second advance, while the hills on the northern and southern sides of Prince Albert Sound may be lateral moraines pushed up along the margins of a lobe that continued to advance farther west. If that is the case, then the Colville Hills are possibly the terminal moraine of ice that readvanced northwards, but did not push far enough into Victoria Island to destroy the old eskers in the center of Wollaston Peninsula.

Apart perhaps from the small northward movement just suggested to account for the Colville Hills, the most recent advance of the ice into Victoria Island seems to have been initiated on the continental mainland south of King William Island. Linear patterns indicate that this ice moved in a general northwesterly direction so as to cross all King William Island and much or all of the lowland in the southern part of Victoria Island. The starting-point of the ice that overran the whole of Victoria Island at an earlier period is not known.

Features visible on aerial photographs suggest local glaciation of the Shaler Mountains on Victoria Island. However, since the rocks here are much disturbed, particularly by faulting, definite proof or disproof of such glaciation may not be forthcoming without detailed ground study.

Mt. Pelly at Cambridge Bay, the conspicuous sand plains north of Washburn Lake and at the head of Richard Collinson Inlet, and certain other isolated features, all seem related in some manner to the former glaciation of Victoria Island. The location of Mt. Pelly between the south and east coasts, both of which were scenes of ice invasion from the continent, arouses the suspicion that it may be a proglacial delta; this suspicion is strengthened by its composition and shape, and by the fact that it seems to have been submerged right to its summit (Washburn, 1947, p. 67).

Several different lines of negative evidence suggest that northern Banks Island, northern Prince of Wales Island, and at least the northern part of Somerset Island, lay beyond the limits of the advancing continental ice. These indications include the absence of lakes, the absence of characteristic glacially deposited material, and the absence of disorganized drainage patterns comparable to those found on the certainly glaciated lands in the southern part of the Archipelago. The fact that the high plateau in northeastern Somerset Island is capped by a uniformly distributed gray material which is

not colored like that of ground moraine farther west strengthens the hypothesis that this plateau may have escaped continental glaciation.

A study of aerial photographs and observations carried out from the air and on the ground at a number of landing points reveal no evidence that the continental glaciation of the Western Arctic extended northward far enough to affect the islands north of Viscount Melville Sound. Parts of that region may have escaped every form of glaciation. However, Nichols' evidence or local glaciation on Cornwallis Island, the cirque-like features on Ellef Ringnes Island, and the certainty of climatic fluctuations in past ages make it reasonable to suppose that many and perhaps most of the highest upland surfaces were once covered by snowfields or ice-caps, and that these cappings resulted in ice-flowage in at least the two places just mentioned. However, since the photographs fail to reveal any further evidence of glaciation, such cappings, like the snowfield that now occupies Meighen Island, may have remained dormant or semi-dormant in many places, perhaps through limited precipitation.

All the recorded movements of continental ice in the Western Arctic Islands are probably related to various phases of the last or Wisconsin glaciation, as it seems unlikely that even the oldest depositional features (some of the eskers on Victoria Island) could have persisted from any earlier glaciation. It is not known whether the Archipelago was ever glaciated prior to Wisconsin time.

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Chart 8C

Series	European Stages (Arkell, 1946)	English Formations (After Arkell 1933)	Northwest Europe Standard Zones (Arkell 1946)	Characteristic fossils in the Western Interior Region
Overlying beds				
UPPER JURASSIC	Portlandian	Purbeck beds		?
		Portland beds	Titanites giganteus	
			Karberites okusensis	
	Kimmeridgian	Kimmeridge clay	Glaucolithites gorei	Vetulonaia sp. and Gyraulus vetarnus
			Zaraiskites albani	
			Pavlovia pallasioides	
			Pavlovia rotunda	
			Pectinatites pectinatus	
			Subplanites wheatleyensis	
			Subplanites sp.	
			Gravesia gigas	
			Gravesia gravesiana	
			Aulacostephanus pseudomutabilis	
			Rasenia mutabilis	
			Rasenia cymodoce	
			Pictonia baylei	
	Oxfordian	Corallian beds	Ringsteadia pseudocordata	?
			Decipia decipiens	
			Perisphinctes coutilsnigroe	
			Perisphinctes plicatilis	
		Oxford clay	Cardioceras cordatum	Cardioceras sp.
			Quenstedtoceras mariae	Cardioceras cordiforme
			Quenstedtoceras lamberti	Quenstedtoceras colliери
	Callovian		Peltoceras athleta	?
			Erymnoceras coronatum	
			Kosmoceras jason	
		Kellaways beds	Sigaloceras calloviense	Kepplerites aclaroni
			Proplanulites koenigi	Kepplerites cf. tychonis
		Cornbrash beds	Macrocephalites macrocephalus	Gowericeras subitum
MIDDLE JURASSIC				Arcticoceras
	Bathonian		Clydoniceras discus	Arctocephalites
		Great oolite	(not yet determined)	
	Bajocian		Parkinsonia parkinsoni	Stemmatoceras and Chondroceras (not zoned)
		Inferior oolite	Stephanoceras humphriesianum	
			Otoites sauzei	
			Sonninia sowerbyi	
			Ludwigia murchisonae	
			Lioceras opalinum	
LOWER JURASSIC	Toarcian	Upper Lias	Lytoceras jurense	
			Hildoceras bifrons	
			Harpoceras serpentinum	
			Dactylioceras tenuicostatum	
		Middle Lias	Paltopleuroceras spinatum	
			Amaltheus margaritatus	
	Pliensbachian	Lower Lias	Prodactylioceras davoei	?
			Tragophylloceras ibex	
			Uptonia jamesoni	
	Sinemurian		Echioceras raricostatum	
			Oxynoticeras oxynotum	
			Asteroceras obtusum	
			Arietites furneri	
			Arnioceras semicostatum	
			Coroniceras bucklandi	
	Hettangian		Scamnoceras angulatum	
			Psiloceras planorbis	
Underlying beds				

	Nevada	Arizona		New Mexico						Colo., Okla., N. M.	Utah									
	Goodsprings Quadrangle and Muddy Mountains, Nevada	Lees Ferry, Arizona	Kayenta, and Rough Rock, Arizona	Beclabito Dome, New Mexico and Red Rock, Arizona	Toadlena and Todilto Park, New Mexico	Window Rock and Lupton, Arizona and Zuni, New Mexico	Fort Wingate Navajo Church, Thoreau and Bluewater, New Mexico	Laguna, El Rito, San Ysidro, Cuchillo Blanco, and Gallina, New Mexico	Navajo Canyon near Ghost Ranch, 6 to 8 miles southeast of Chama, New Mexico	Gallinas River Area S.E. of Las Vegas and Ima Area S.W. of Tucumcari, N. M.	Gallegos Area, N.M. and Cimarron Rim Area of Colorado, Oklahoma and New Mexico	Iron Springs, Diamond Valley, Zion Canyon, Kanab Canyon, Utah	Paria Canyon, Utah	Navajo Mountain, Utah	Circle Cliffs, Indian Creek, Bluff and Comb Ridge, Boundary Butte, Utah	Green River Desert, Utah	San Rafael Swell at Starvation Cr., Saddle Horse Can., Horn Silver Gulch, Black Dragon Can., Utah	Mouth of San Rafael Riv., Salt Valley, Dewey Bip Hole, Sinbad Ranch, Utah	Wasatch Plateau-Saline to Indianola, Central Utah	Wasatch Mts., Thistle to Emigration Canyon, Utah	Uinta M., Provo Pass to Fork,
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	2
Overlying beds	Tertiary / Upper Cretaceous		Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous		Creta
	?	?	?	?	?	?	?	?	?	?	?	?		?	?	?	?	?	?	?	?
			Morrison formation 1	Morrison formation 1	Morrison formation 1	Zuni sandstone 1	Morrison formation 1	Morrison formation 1	Morrison formation 1	Morrison formation 1	Morrison formation 1			Morrison formation 1	Morrison formation (Members apply to Blanding-Comb Ridge) 1	Morrison formation 1	Morrison formation 1	Morrison formation 1	Morrison (?) formation 10	Morrison (?) formation 10	Morri forma
					Morrison formation 1		Morrison formation 1								Brushy Basin shale member						
															Westwater Canyon sandstone member						
															Recapture shale member						
					Zuni sandstone		Zuni sandstone								Bluff sandstone member						
			?	?	?		?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
			Wanakah formation 2	Sandstone and shale 2	Wanakah formation 2	Sand-stone	Wanakah formation 2	Wanakah formation 2	Wanakah formation 2	Wanakah formation 2	Wanakah formation 2		Winsor formation 6			Summerville formation 9	Summerville formation 9	Summerville formation 9			
			Todilto limestone member	Todilto limestone member	Todilto limestone member		Todilto limestone member	Todilto limestone member (Gypsiferous facies)	Todilto limestone member	Todilto limestone member	Todilto limestone member (Gypsiferous facies)	Curtis formation 7	Curtis formation 7		Summerville formation 9	Summerville formation 9	Curtis formation 7	Curtis formation 7	Moab tongue 3	Curtis formation 11	Curtis formation 7	Stun sandst
			?	?	?		?	?	?	?	?	?	?		?	?	?	?	?	?	?
			Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Entrada ss. (Exp'or co.) 3	Entrada sandstone (Lower part of Plateau co.) 3	Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Entrada sandstone 3	Twist Gulch member 11	Prouss sandstone 13	Preus sandst
			?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
												Carmel formation 4									
			Carmel formation 4	Carmel formation 4	Carmel formation 4	Carmel formation 4	Probably present 4	Carmel formation 4	Carmel formation 4	Carmel formation 4		(Equals Homestake limestone of Iron Springs District)	Carmel formation 4	Carmel formation 4	Carmel formation 4	Carmel formation 4	Carmel formation 4	Carmel formation 4	Arapien shale 11	Twin Creek limestone 14	Twin Creek limestone 14
			?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
			Navajo sandstone 5	Navajo sandstone 5	Navajo sandstone 5							Navajo sandstone 5	Navajo sandstone 5	Navajo sandstone 5	Navajo sandstone 5	Navajo sandstone 5	Navajo sandstone 5	Navajo sandstone 5	Not exposed	Nugget sandstone 15	Nugget sandston
			"Aztec sandstone"																		
			?	?	?							?	?	?	?	?	?	?		?	?
			Kayenta formation / Wingate sandstone	Kayenta fm. / Wingate sandstone	Absent at Red Rock	Glen Canyon group undivided 6	Glen Canyon group undivided 6	Glen Canyon group undivided 6				Kayenta formation / Wingate sandstone		Kayenta formation / Wingate sandstone	Kayenta formation / Wingate sandstone	Kayenta formation / Wingate sandstone	Kayenta formation / Wingate sandstone	Kayenta formation / Wingate sandstone			
Underlying beds	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic		Triassic	Triassic

Chart 8C.- CORRELATION OF THE JURASSIC ROCKS OF THE WESTERN INTERIOR OF THE UNITED STATE

By Ralph W. Imlay

	Utah									Utah		Colorado										Colorado									
	Iron Springs, Diamond Valley, Zion Canyon, Kanab Canyon, Utah	Paria Canyon, Utah	Navajo Mountain, Utah	Circle Cliffs, Indian Creek, Bluff and Comb Ridge, Boundary Butte, Utah	Green River Desert, Utah	San Rafael Swell at Starvation Crk., Saddle Horse Can., Horn Silver Gulch, Black Dragon Can., Utah	Mouth of San Rafael Riv., Salt Valley, Davey, Big Hole, Sid Scarf Ranch, Utah	Wasatch Plateau-Saline to Indianola, Central Utah	Wasatch Mtns., Thistle to Emigration Canyon, Utah	Uinta Mountains Poco to Lake Fork, Utah	Uinta Mountains White Rocks to Split Mountain, Utah	Skull Creek and Vermilion Creek, Colorado	Western and southern margins of San Juan Mountains, Colorado	Bush Canyon, Gypsum Valley, and Dry Creek, Colorado	Sinbad Valley and Bedrock, Colorado	Serpents Trail and Unaweep Canyon, Colorado	Bostwick Park, Colorado	Sapinero and Gunnison, Colorado	S.E. of Meeker, Rifle Creek, South Canyon, Crystal Riv., Scofield Park, Colorado	Phippsburg, Burns, Wolcott, Snowmass Can., Colorado	Hahns Peak and North Park Areas, Colorado	Breckenridge and Montezuma Areas, Colorado	Boulder Creek to Boulder, Colorado	Ralston Creek to Morrison, Colorado	Turkey Creek to Eight Mile, Colorado	Canyon City and North Creek, Colorado	Beulah to Vermejo Creek, Colorado	Pur... Sou... Co...
	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	
	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous		Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cre...
	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	
	7 Curtis formation	8 Winsor formation	1 Morrison formation	1 Morrison formation (Members apply to Bluff and Comb Ridge) Brushy Basin shale member; Westwater Canyon sandstone member; Recapture shale member; Bluff sandstone member	1 Morrison formation; Salt Wash sandstone member	1 Morrison formation; Salt Wash sandstone member	1 Morrison formation; Salt Wash sandstone member	10 Morrison (?) formation	10 Morrison (?) formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	
	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	
	9 Summerville formation	7 Curtis formation	9 Summerville formation	9 Summerville formation	7 Curtis formation	7 Curtis formation	3 Moab tongue	11 Curtis formation	7 Curtis formation	12 Stump sandstone	7 Curtis formation	7 Curtis formation	2 Junction Creek sandstone member; Morrison formation: Tidwell member, Salt Wash member, Recapture member, Westwater Canyon member, Brushy Basin member; Wanakah formation: Pony Express limestone member	9 Summerville formation	9 Summerville formation	9 Summerville formation	2 Wanakah (?) formation		7 Curtis formation	7 Curtis fm.									
	3 Entrada sandstone (Lower part of Pinto ss.)	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	11 Twist Gulch member	13 Preuss sandstone	13 Preuss sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	?	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada sandstone	3 Entrada ss.	3 Entrada ss.	3 Entrada ss.	3 Entrada sandstone	3 Entrada sandstone	
	4 Carmel formation (Equals Homestake limestone of Iron Springs District)	4 Carmel formation	4 Carmel formation	4 Carmel formation	4 Carmel formation	4 Carmel formation	4 Carmel formation	11 Arapien shale	14 Twin Creek limestone: Shale and soft shaly limestone, light gray; Limestone, thin to thin-bedded, oolitic in part; Red and green siltstone and yellow silty limestone; Limestone, shaly, soft, light gray; Ls., medium to thin-bedded, partly sandy, oolitic at base, brownish; Red and green siltstone and brecciated limestone	14 Twin Creek limestone: Ls., sandy; Limestone, shaly, soft light gray; Ls., medium to thin bedded, some oolitic, hard; Red beds and yellow, silty, sandy limestone, ripple marked; Limestone, shaly, soft, light gray; Ls., medium to thin-bedded, smooth, yellowish gray	4 Carmel formation		4 Carmel formation																	
	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	
	5 Navajo sandstone	5 Navajo sandstone	5 Navajo sandstone	5 Navajo sandstone	5 Navajo sandstone	5 Navajo sandstone	5 Navajo sandstone	Not exposed	15 Nugget sandstone	15 Nugget sandstone	15 Nugget sandstone	15 Nugget sandstone	Wingate sandstone	5 Navajo sandstone	5 Navajo sandstone	Kayenta to Wingate sandstone	Wingate sandstone											
	?	?	? Kayenta formation; Wingate sandstone	? Kayenta formation; Wingate sandstone	? Kayenta formation; Wingate sandstone	? Kayenta formation; Wingate sandstone	? Kayenta formation; Wingate sandstone		?	?	?	?		? Kayenta formation; Wingate sandstone	? Kayenta formation; Wingate sandstone													
	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic		Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Pre-Cambrian	Pre-Cambrian	Triassic / Carboniferous	Triassic	Triassic	Permian	Triassic	Permian	Permian	Permian	Permian / Pennsylvanian	

THE WESTERN INTERIOR OF THE UNITED STATES

W. Imlay

Colorado										Kansas	Idaho and Wyo.	Wyoming										So. Dak. and Wyo.	Nebraska	North Dakota	
S.E. of Meeker, Rifle Creek, South Canyon, Crystal Riv., Scofield Park, Colorado	Phippsburg, Burns, Wolcott, Snowmass Can., Colorado	Hahns Peak and North Park Areas, Colorado	Breckenridge and Montezuma Areas, Colorado	Boulder Creek to Boulder, Colorado	Ralston Creek to Morrison, Colorado	Turkey Creek to Eight Mile, Colorado	Canyon City and North Creek, Colorado	Beulah to Varmeja Creek, Colorado	Purgatoire Uplift, Southeastern Colorado	Western Kansas	Southeastern Idaho and southwestern Wyoming	Hogback and Teton Mountains, Jackson Hole Area, Wyoming	South side Yellowstone Park and Gros Ventre Mtns., Wyo.	Green River Lakes Area at west end Wind River Mountains., Wyoming	Dubois Area and south side of Owl Creek Mountains., Wyoming	Wind River Basin, southwest part from Lander to Conant Creek and Bison Basin, Wyoming	West of southeastern Wyoming east of Dutton Basin and south of Powder River, Wyoming	Southeastern Wyo.-Colo. border areas, including Horse Cr., Centennial Valley and Sierra Madre, Wyo.	Rock Springs Area, or Baxter Basin	Cody Area on west side of Big Horn Basin, Wyoming	Big Horn Mtns. except southeast end, Wyoming	Southeast end of Big Horn Mtns. and Salt Creek Area, Wyoming	Black Hills, northeast Wyo., and western South Dakota	Western Nebraska	Western North Dakota	West-central and northwestern Montana
30	31	32	33	34	35	36	37	38	39	40	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56
Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous
1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	17 12 Stump sandstone	17 12 Stump sandstone	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation / Unkpapa sandstone (Southeast and south sides only)	1 Morrison formation	1 Morrison formation	1 Morrison formation (Absent locally at north end of Sweetgrass Arch)
7 Curtis formation	7 Curtis formation	7 Curtis fm.				Wanakah formation / Todilto ls. member			Wanakah formation / Todilto limestone member (present)				12 Stump sandstone	22 "Upper Sundance" formation	22 "Upper Sundance" formation	22 "Upper Sundance" formation	22 "Upper Sundance" formation	22 "Upper Sundance" formation	7 Curtis formation	23 Glauconitic sandstone and shale / Conglomeratic at base locally	23 Glauconitic shale and sandstone	23 Glauconitic shale and sandstone	23 Redwater shale member / Lak member / Hulett ss. member / Skiboda Beaver shale member / Canyon Springs sandstone member	23 Sundance formation (Thins out eastward)	23 Grossish gray shale, some sandstone near top / Pink to red-brown sh. / Gray shale and fine ss. / Gray shale some limestone and sandstone	24 Swift formation / 25 Rierdon formation (Absent on South Arch and Big Belt Uplift) / 26 Sawtooth formation (Pinches out on South Arch and Big Belt)
Entrada sandstone 3	Entrada sandstone 3		Entrada ss. 3	3 Entrada sandstone	Entrada ss. 3	Entrada ss. 3			3 Entrada sandstone	3 Entrada sandstone	13 Preuss sandstone	13 Preuss sandstone							3 Entrada sandstone	3 Entrada sandstone						
											14 Ls., sandy / Limestone, shaly / Ls., medium thin-bedded, some oolite / Red beds and yellow silty-sandy limestone / Limestone, shaly, medium gray / Ls., black, medium bedded, oolitic at base some ss. and conglomerate / Brecciated red beds, limestone and dolomite	14 Ls., sandy / Limestone, shaly / Ls., medium thin-bedded, oolitic at base / Red beds / Dark shale and limestone oolitic at base / Brecciated red beds, limestone and dolomite														
Twin Creek limestone													14 Gray shale with oolitic limestone at base / Red beds / Dark shale and limestone, oolitic at base / Brecciated red beds, and limestone	18 Gray shale with oolitic limestone at base / Red beds / Dark shale and oolitic limestone	18 Gray ss. / Gray shale / Limestone, oolitic shale, siltstone and limestone / Red beds at top gray limestone and shale at base	18 Lak member / Hulett ss. member / Skiboda Beaver shale member / Canyon Springs sandstone member	18 Lak member / Hulett ss. member / Skiboda Beaver shale member / Canyon Springs ss.	Entrada (?) 3 sandstone (Probably present locally)	3 Entrada sandstone / 4 Carmel formation						
														19 Gypsum Spring formation	19 Gypsum Spring formation	19 Gypsum Spring formation				20 Red sh. and siltstone / Gray shale, limestone, dolomite, and thin beds of gypsum / Gypsum, dolomite and red beds	20 Gypsum Spring formation of authors	Gypsum Spring fm.			
												15 Nugget sandstone	15 Nugget sandstone	15 Nugget sandstone	15 Nugget sandstone (Absent on north side of Owl Creek Mountains)	15 Nugget sandstone	15 Nugget sandstone (Probably absent locally)	15 Nugget sandstone (Probably absent locally)	15 Nugget sandstone				18 Nugget (?) sandstone (Southwest end only)	18 Nugget (?) sandstone (Northwest end of state)	21 Piper formation	
Triassic / Carboniferous	Triassic	Triassic	Permian	Triassic	Permian	Permian	Permian	Permian / Pennsylvanian	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic to Permian	Triassic to Devonian	Pennsylvanian to Mississippian	Penn. to Mis.

TES

Prepared under the auspices of the Committee on Stratigraphy of the National Research Council, Carl O. Dunbar, *Chairman*

	Kansas	Idaho and Wyo.	Wyoming											So. Dak. and Wyo.	Nebraska	North Dakota	Montana								
Purgatoire Uplift, Southeastern Colorado	Western Kansas	Southeastern Idaho and southwestern Wyoming	Hogback and Teton Mountains, Jackson Hole Area, Wyoming	South side Yellowstone Park and Gros Ventre Mtns., Wyo.	Green River Lakes Area at west end Wind River Mountains., Wyoming	Dubois Area and south side of Owl Creek Mountains., Wyoming	Wind River Basin, southwest part from Lander to Conant Creek and Bison Basin, Wyoming	Most of southeastern Wyoming east of Dutton Basin and south of Powder River, Wyoming	Southeastern Wyo.-Colo. border areas, including Horse Crk., Centennial Valley and Sierra Madre, Wyo.	Rock Springs Area, or Baxter Basin	Cody Area on west side of Big Horn Basin, Wyoming	Big Horn Mtns. except southeast end, Wyoming	Southeast end of Big Horn Mtns. and Salt Creek Area, Wyoming	Black Hills, northeast Wyo., and western South Dakota	Western Nebraska	Western North Dakota	West-central and northwestern Montana	North-central Montana east of Sweetgrass Arch	North flank of Little Belt Mountains	South flank of Little Belt Mountains	North flank of Big Snowy Mountains	South flank of Big Snowy Mountains	Southwestern Montana Madison Range, Three Fork–Cardwell Area	Southern Montana Pryor Mountains to Gallatin Range	Eastern Montana (subsurface)
39	40	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	62	63	64
Cretaceous	Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous
1 Morrison formation	1 Morrison formation	17 1 Morrison formation	17 1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation / Unkpapa sandstone (Southeast and south sides only)	1 Morrison formation	1 Morrison formation	1 Morrison formation (Absent locally at north end of Sweetgrass Arch)	(Absent in Sweetgrass Hills)	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation	1 Morrison formation
2 Wanakah formation (present) / Todilto limestone member		12 "Stump" sandstone	12 Stump sandstone	12 Stump sandstone	22 "Upper Sundance" formation	22 "Upper Sundance" formation	22 "Upper Sundance" formation	22 "Upper Sundance" formation	22 "Upper Sundance" formation	7 Curtis formation	23 Sundance formation: Glauconitic sandstone and shale; Conglomerate at base locally	23 Sundance formation: Glauconitic shale and sandstone	23 Sundance formation: Glauconitic shale and sandstone	23 Sundance formation: Redwater shale member	23 Sundance formation (Thins out westward)	23 Sundance formation: Greenish gray shale, some sandstone near top	24 Swift formation	24 Swift formation	24 Swift formation	24 Swift formation	24 Swift formation	24 Swift formation	24 Swift formation	24 Swift formation	24 Swift formation
3 Entrada sandstone		13 Preuss sandstone	13 Preuss sandstone	14 Gray shale with oolitic limestone at base; Red beds; Dark shale and limestone, oolitic at base; Brecciated red beds, limestone and dolomite	18 Gray shale with oolitic limestone at base; Red beds; Dark shale and oolitic limestone; 19 Gypsum Spring formation	18 Gray shale; Limestone, oolitic shale, siltstone and limestone; Red beds at top gray limestone and shale at base; 19 Gypsum Spring formation	18 Lak member; Hulett ss. member; Hulett sandstone shale member; Canyon Springs sandstone member; 19 Gypsum Spring formation	18 Lak member; Hulett ss. member; Stokoda Beaver shale member; Canyon Springs ss.; 19 Gypsum Spring formation	3 Entrada (?) sandstone (Probably present locally)	3 Entrada sandstone	20 Gypsum Spring formation of authors: Red sh. and siltstone; Gray shale, limestone, dolomite, and thin beds of gypsum; Gypsum, dolomite and red beds	20 Gypsum Spring formation of authors	20 Gypsum Spring formation of authors	19 Gypsum Spring fm.	23 Sundance formation: Pink to red-brown sh.; Gray shale and fine ss.; Gray shale some limestone and sandstone	25 Ellis group: Rierdon formation (Absent on South Arch and Big Belt Uplift)	25 Rierdon formation	25 Rierdon formation	25 Rierdon formation	25 Rierdon formation (Absent at east and west ends of mountains)	25 Rierdon formation	25 Rierdon formation (Thins westward)	25 Rierdon formation (Thins westward and northward)	25 Rierdon formation
Twin Creek limestone: 14 Limestone, sandy; Limestone, shaly; Ls., medium thin-bedded, some oolitic; Red beds and yellow silty-sandy limestone; Limestone, shaly, medium gray; Ls., black, medium bedded oolitic at base some ss. and conglomerate; Brecciated red beds, limestone and dolomite			14 Limestone, sandy; Limestone, shaly; Ls., medium thin-bedded, oolitic at base; Red beds; Dark shale and limestone oolitic at base; Brecciated red beds, limestone and dolomite	14 Twin Creek limestone																						
			15 Nugget sandstone		15 Nugget sandstone	15 Nugget sandstone (Absent on north side of Owl Creek Mountains)	15 Nugget sandstone	15 Nugget sandstone (Probably absent locally)	15 Nugget sandstone (Probably absent locally)	15 Nugget sandstone				15 Nugget (?) sandstone (Southwest end only)	15 Nugget (?) sandstone (Northwest end of state)	21 Piper formation	26 Sawtooth formation (Pinches out on South Arch and Big Belt)	21 Piper formation (Absent in Little Rocky Mountains)	21 Piper formation (Overlap westward and southward)		21 Piper formation (Pinches out southward)		26 Sawtooth formation (Pinches out northward)	21 Piper formation (Pinches out northward)	21 Piper formation
Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic	Triassic to Permian	Triassic to Devonian	Pennsylvanian to Mississippian	Pennsylvanian to Mississippian	Pennsylvanian to Mississippian	Pennsylvanian	Pennsylvanian to Mississippian	Triassic to Mississippian	Triassic to Permian	Triassic to Pennsylvanian	Triassic to Mississippian

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VOL. 53, PP. 953-992 SEPTEMBER 1942

CORRELATION OF THE JURASSIC FORMATIONS OF NORTH AMERICA, EXCLUSIVE OF CANADA

By Ralph W. Imlay

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Introduction

This report on the Jurassic formations of North America (exclusive of Canada) is Number 8C of a series of correlation charts prepared under the auspices of the Committee on Stratigraphy of the National Research Council.[1] Under the sponsorship of the U. S. Geological Survey, the writer spent approximately 1 year in field and office gathering and compiling information pertinent to it. For convenience of discussion the Jurassic formations are grouped under five large regions, namely, the Gulf region, the Western Interior of the United States, the Pacific Coast region of the United States, Alaska, and Greenland. For the convenience of future workers fairly complete lists of references are given separately for each region. The large number of columns devoted to the Western Interior of the United States merely reflects the greater amount of field work done there than in the other regions and does not mean that the Jurassic stratigraphy of the Western Interior is more complex than elsewhere. The chart is intended to summarize existing knowledge concerning the Jurassic of North America (excepting Canada) and to indicate the areas or problems where more detailed study is most needed. The Jurassic formations of western Canada will be treated by Dr. Hans Frebold in Chart 8D.

Age equivalencies are shown by horizontal lining, hiatuses by vertical lining, and lack of knowledge by oblique lining. The annotation are numbered to agree with numerals on the chart.

Index maps show localities mentioned in this report or commonly referred to in publications. No attempt has been made to show the positions of all well-exposed sections as that would require many additional maps and is not necessary for the purposes of this paper.

Annotations

1. The Upper Jurassic age of the Morrison formation is now accepted without question (Baker, Dane, and Reeside, 1936, p. 58–63), but its exact upper age limit is not known. At least part of the formation must be of Kim-

[1] Dunbar, C. O., *et al.* (1942) *Correlation charts prepared by the Committee on Stratigraphy of the National Research Council*, Geol. Soc. Am., Bull., vol. 53, p. 429–434.

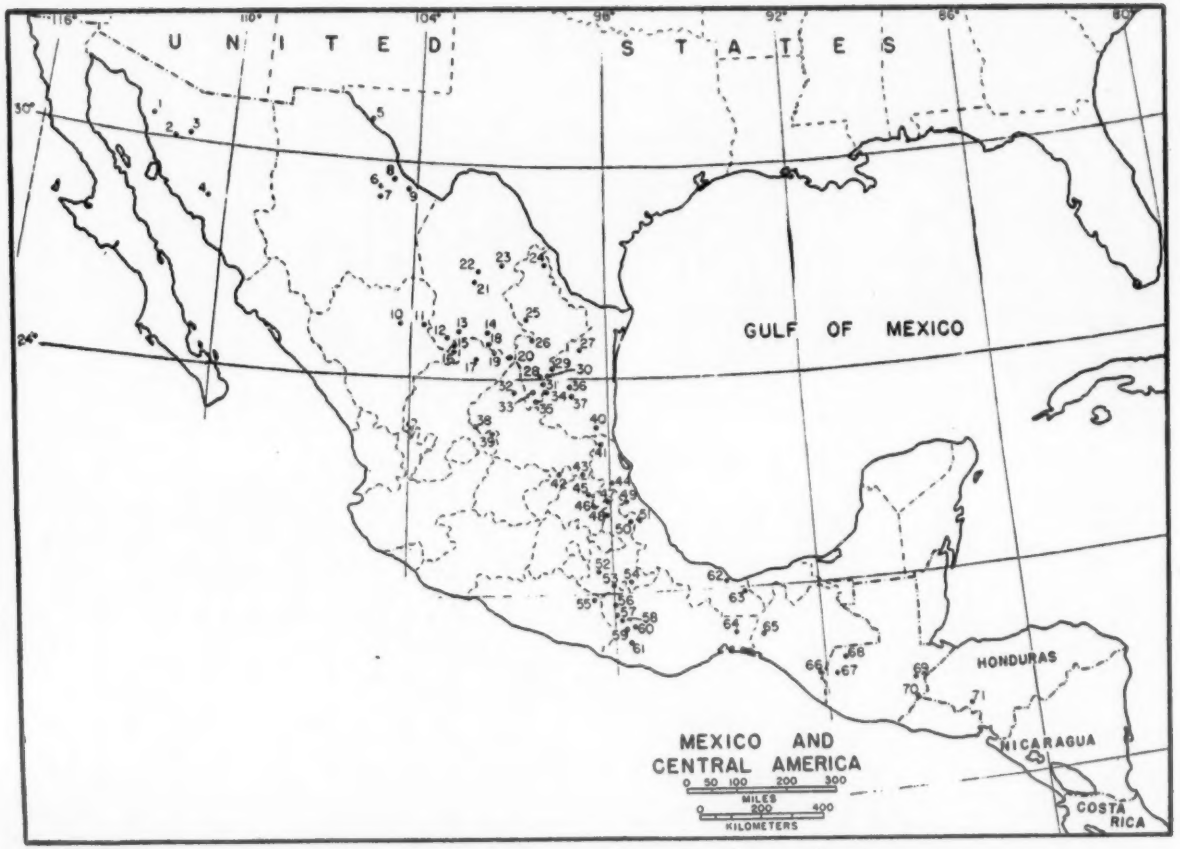

FIGURE 1.—JURASSIC LOCALITIES IN MEXICO AND CENTRAL AMERICA

1. El Antimonio
2. Sierra de Santa Rosa
25. Sierras between Monterrey and Saltillo
26. Cienega del Toro
49. Poza Rica Well No. 8
50. Tejiutlan

FIGURE 1.—JURASSIC LOCALITIES IN MEXICO AND CENTRAL AMERICA

1. El Antimonio
2. Sierra de Santa Rosa
3. Las Animas
4. Sierra de Moradillas
5. Malone
6. Placer de Guadalupe
7. Los Fresnos del Río
8. Cuchillo Parado
9. Sierra Rica
10. San Pedro del Gallo
11. Villa Juarez
12. Sierra de Jimulco
13. Western part of Sierra de Parras
14. Middle part of Sierra de Parras
15. Sierra de Ramirez
16. Sierra del Chivo
17. Pico de Teyra
18. Sierras de Zuloaga and Sombreretillo
19. Sierras near Mazapil
20. Sierra de la Ventura
21. San Marcos
22. Menchaca
23. Sierra Azul
24. San Ambrosio No. 1 well
25. Sierras between Monterrey and Saltillo
26. Cienega del Toro
27. Northeastern part of Sierra de San Carlos
28. La Escondida
29. Las Virgenes
30. Mesquital
31. San Lazaro
32. Sierra de Catorce
33. Doctor Arroyo
34. Miquihuana
35. Mier y Noriega
36. Sierra west of Ciudad Victoria
37. Huizachal
38. Peñon Blanco
39. Pinos
40. Chocoy Well No. 2 and Altamira Well No. 11
41. Panuco Well No. 82
42. Jalpan
43. Tamazunchale
44. Tlacolula No. 8 well
45. Puerto de Tlamacho
46. Huayacocotla
47. Huehuetla
48. Huauchinango
49. Poza Rica Well No. 8
50. Teziutlan
51. Almanza
52. Matamoros
53. Acatlan
54. Zapotitlan
55. Cualac
56. Consuelo
57. Mixtepec
58. Tlaxiaco
59. Santo Domingo Ixcatlan
60. Rio Niochistlan north of Yutanduchi
61. Amoltepec
62. Chinameca
63. Rio Playas
64. Rincon Antonio-Magoñe area
65. Southeast of Cintalapa
66. Motocintla
67. Todos Santos
68. Quetzal
69. Jocatan
70. Metapan
71. Tegucigalpa

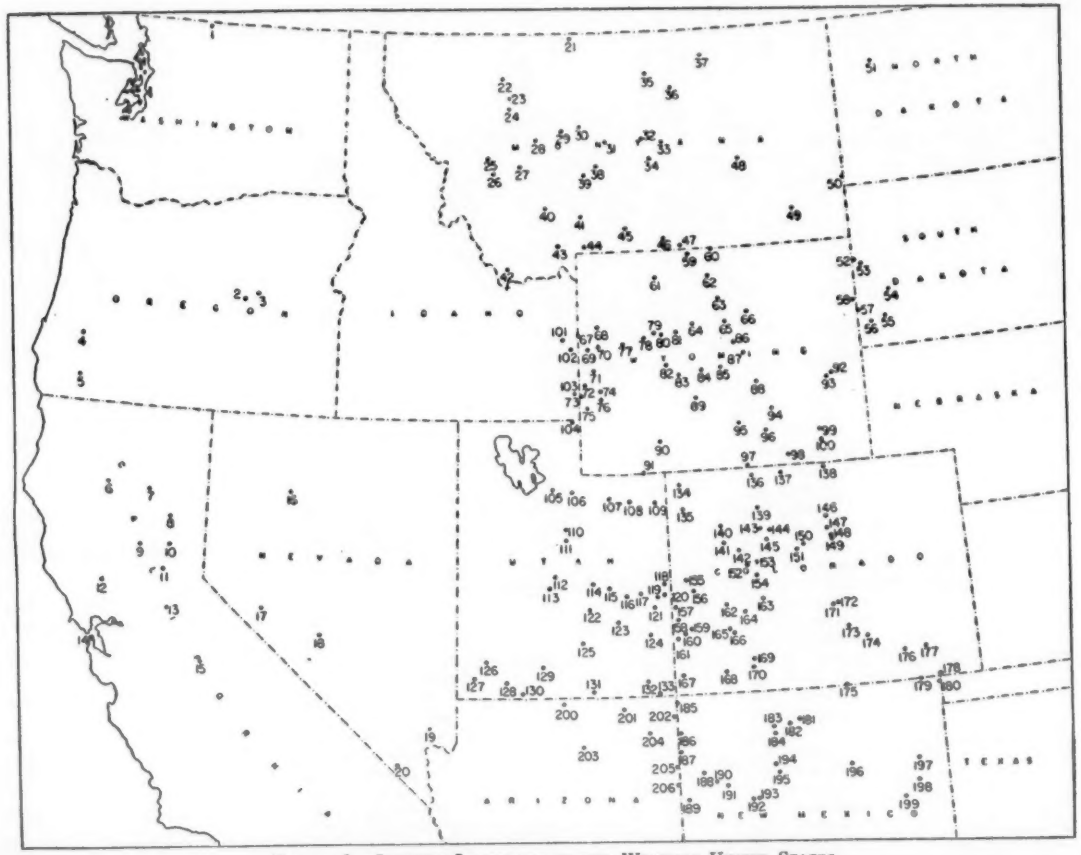

FIGURE 2.—JURASSIC LOCALITIES IN THE WESTERN UNITED STATES

1. North end of Cascade Range
2. South Fork Valley
43. Indian Creek, Madison Range
44. Cinnabar Mountain
78. Red Grade
79. Red Creek
80. Maverick Springs
121. Wilson Mesa
122. Starvation Creek
123. Green River Desert
166. Ouray
167. McElmo Canyon
168. Durango

FIGURE 2.—JURASSIC LOCALITIES IN THE WESTERN UNITED STATES

1. North end of Cascade Range
2. South Fork Valley
3. Head of Silvies River
4. Roseburg Area
5. Grants Pass Area
6. Redding
7. Lassen Park
8. Taylorsville
9. Oroville
10. Downieville
11. Colfax
12. Knoxville
13. Amador Area
14. San Francisco
15. Mariposa
16. Humboldt Range
17. Hawthorne
18. Tonopah
19. Muddy Mountains
20. Goodsprings Quadrangle
21. East Butte of Sweetgrass Hills
22. Swift Reservoir
23. Rierdon Gulch
24. Sun River
25. Drummond Area
26. Gird Creek
27. Elliston Area
28. Missouri Canyon
29. Smith River
30. Belt Creek
31. Three miles S.W. of Utica
32. Near Heath
33. Button Butte Dome
34. Stone House Ranch
35. Section Creek
36. Chalk Butte Dome
37. The Texas Co.—Bowdoin Unit 1
38. Delpine
39. Moss Agate Area
40. Sappington Canyon
41. Rocky Creek Canyon
42. Little Water Canyon
43. Indian Creek, Madison Range
44. Cinnabar Mountain
45. Benbow Mill Road
46. Red Dome
47. Gypsum Creek
48. Prairie—Westphal No. 1-XX
49. Stanolind—Northern Pacific No. 1
50. Montana - Dakota - Northern Pacific No. 1
51. California—Kamp No. 1
52. Redwater Creek
53. Near Spearfish
54. Six miles south of Rapid Creek
55. Two miles N.W. of Buffalo Gap
56. Three miles N.W. of Minnekahta
57. East side of Elk Mountain
58. Stockade Beaver Creek
59. North of Kane
60. Red Gulch
61. Near Cody
62. Near Shell
63. Five miles N.W. of Tensleep
64. Thermopolis
65. North of Nowood
66. Mayoworth
67. Mosquito Pass
68. Gros Ventre Slide
69. Cabin Creek
70. Near Camp Davis
71. Sheep Creek at Greys River
72. Cottonwood Creek
73. Thomas Fork
74. South Piney Creek
75. Sliderock Creek
76. La Barge Creek
77. Green River Lakes Area
78. Red Grade
79. Red Creek
80. Maverick Springs
81. East fork of Sheep Creek
82. Mill Creek
83. Dallas Anticline
84. Conant Creek
85. Dutton Basin
86. North of Arminto
87. Powder River
88. Alcova
89. Bison Basin
90. Baxter Basin
91. Flaming Gorge
92. Spanish Diggings
93. Near Glendo
94. Difficulty Canyon
95. Near Rawlins
96. Elk Mountain
97. Sierra Madre
98. Centennial Valley
99. Iron Mountain
100. Horse Creek
101. Fall Creek
102. Big Elk Mountain
103. Preuss Creek
104. East of Bear Lake
105. Fort Douglas Area
106. Weber River
107. Lake Fork
108. White Rocks
109. Split Rock (Dinosaur Quarry)
110. Thistle
111. Indianola
112. Gunnison
113. Salina
114. Saddle Horse Canyon
115. Black Dragon Canyon
116. Mouth of San Rafael River
117. Salt Valley
118. Big Hole
119. Dewey
120. Scharf Ranch
121. Wilson Mesa
122. Starvation Creek
123. Green River Desert
124. Indian Creek, Utah
125. Circle Cliffs
126. Iron Springs
127. Diamond Valley
128. Zion Canyon
129. Paria River
130. Kanab Canyon
131. Navajo Mountain
132. Bluff and Comb Ridge
133. Boundary Butte
134. Vermilion Creek
135. Skull Creek
136. Hahns Peak
137. North Park
138. Boxelder Creek
139. Phippsburg
140. S.E. of Meeker
141. Rifle Creek
142. South Canyon
143. Burns
144. State Bridge
145. Wolcott
146. Boulder
147. Ralston Creek
148. Morrison
149. Turkey Creek
150. Montezuma
151. Breckenridge
152. Crystal River
153. Snowmass Canyon
154. Scofield Park
155. Serpents Trail
156. Unaweep Canyon
157. Sinbad Valley
158. Bedrock
159. Dry Creek
160. Gypsum Valley
161. Bush Canyon
162. Bostwick Park
163. Near Gunnison
164. Sapinero
165. Ridgway
166. Ouray
167. McElmo Canyon
168. Durango
169. Piedra River at Wimi-nuche Creek
170. Piedra River at First Fork
171. West of Canyon City
172. Eight Mile Creek
173. Beulah
174. Cuchara Canyon
175. Vermejo Creek
176. Purgatoire River
177. Johnny Creek
178. Carrizo Creek
179. Baker Ranch
180. Cimarron River at State Line
181. El Rito
182. Navajo Canyon near Ghost Ranch
183. Gallinas River
184. Gallina
185. Beclabito
186. Toadlena
187. Todilto Park
188. Fort Wingate and Navajo Church
189. Zuni
190. Thoreau
191. Bluewater
192. Laguna
193. El Rito Siding
194. Cuchillo Blanco
195. San Ysidro
196. S.E. of Las Vegas
197. Gallegos
198. Tucumcari
199. Ima
200. Lees Ferry
201. Kayenta
202. Redrock
203. Tuba
204. Rough Rock
205. Window Rock
206. Lupton

meridgian age, as the underlying marine beds contain ammonites of Oxfordian age (Reeside, 1919; Imlay, 1947, p. 260–261) and lack the pelecypod *Aucella*, which was abundant in

is essentially of Kimmeridgian age and that its highest beds are older than the Purbeck. Perhaps the end of Morrison deposition coincided with the lower Portlandian uplift in the north-

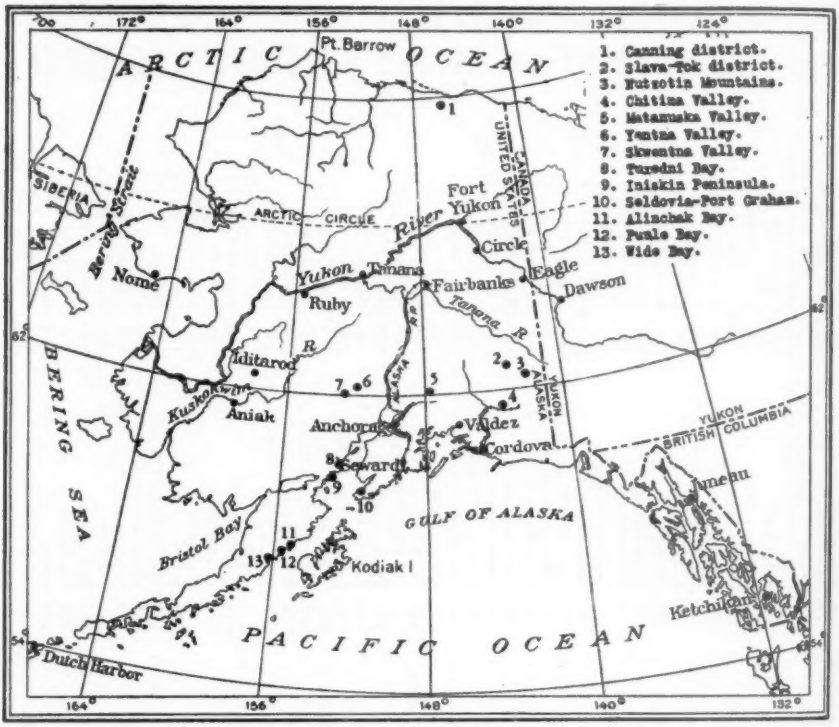

FIGURE 3.—JURASSIC LOCALITIES IN ALASKA

boreal waters during the Kimmeridgian. Its reptiles show close relationships with forms in Portlandian beds from East Africa (Schuchert, 1918; Simpson, 1926) and the mammals and ostracodes with forms in the Purbeck of England (Simpson, 1929; Jones, 1886). However, recent studies by Teng-Chien Yen (Personal communication) show that the molluscan fauna of the Morrison is quite distinct from that in the Purbeck and is probably somewhat older. This checks with the work of Arkell (1941, p. 127, 128), who notes greater resemblances of the gastropods in the Purbeck beds to those in the Lower Cretaceous Bear River formation than to those in the Morrison. These considerations indicate that the Morrison

ern part of the Gulf region, when great thicknesses of conglomeratic sandstone were deposited in Arkansas, Louisiana, Texas, and elsewhere (Imlay, 1943, p. 1490, 1526, 1527; Swain, 1944, p. 597–600, 611, 612). Of course, the unconformity between the Morrison and Cloverly formations in Wyoming, or the Morrison and Kootenai formations in Montana, involves much more than Purbeck or Portlandian time, as shown by the distinctness of the fresh-water molluscan assemblages (Yen, 1946, p. 4; Yen and Reeside, 1946, p. 53), probably by the differences in the microfossils (Peck, 1937; 1941), and by evidence of extensive post-Morrison erosion in western Montana. Concerning the latter, Cobban (1945, p. 1270,

1278, 1282–1290) has noted that after deposition of the Morrison formation the Sweetgrass arch area was elevated, gently folded, and then subjected to erosion that locally removed all of the Morrison and part of the Ellis group before deposition of the lower Cretaceous Kootenai formation. In southwestern Montana, Imlay has noted that the entire Jurassic system, totalling about 600 feet, terminates abruptly on the west in the Madison Range, but that the Kootenai formation continues westward to the Pioneer Range, where it is at least 1500 feet thick. These relationships in Montana should be reflected farther south. Perhaps the variable thickness of the Morrison formation in central Wyoming may be partly a result of post-Morrison erosion.

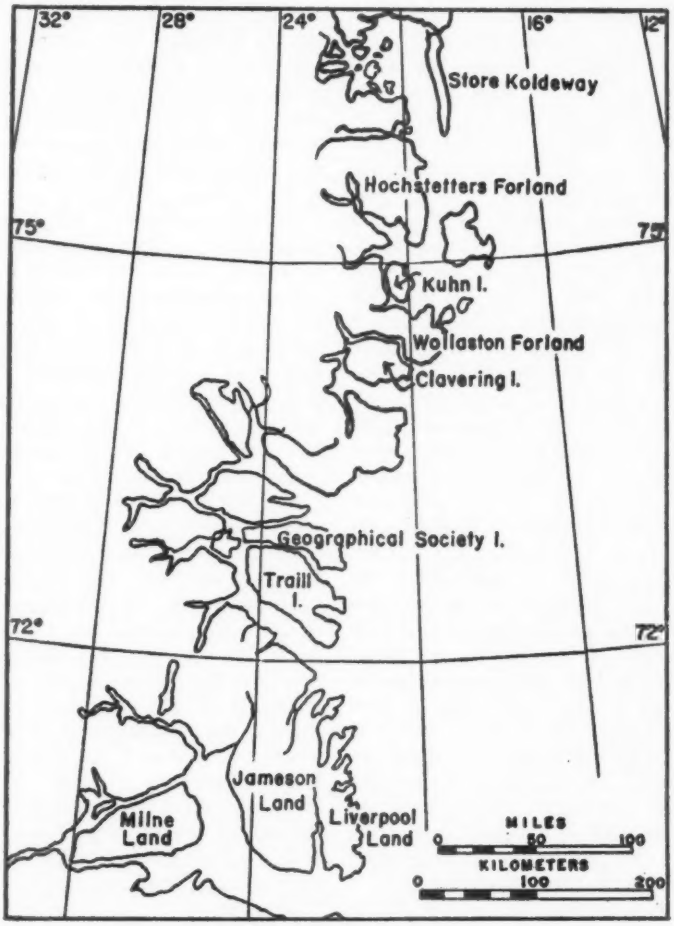

FIGURE 4.—JURASSIC LOCALITIES IN GREENLAND

In some areas the Morrison formation has been subdivided into members that may be useful locally (Silver, 1948, p. 77–79). The Zuni sandstone is as conspicuous in parts of northwestern New Mexico and northeastern Arizona (Dutton, 1885, p. 135–138) as the Unkpapa sandstone is along the southeastern

side of the Black Hills in South Dakota (Darton, 1899, p. 393; 1909, p. 37; Ballard, 1942, p. 1562; Imlay, 1947, p. 246, 247). In parts of east-central Utah the conglomeratic Salt Wash sandstone member characterizes the lower part of the Morrison (Lupton, 1914, p. 127; Baker, Dane, and Reeside, 1936, p. 9, 15; Gilluly and Reeside, 1928, p. 81, 82; McKnight, 1940, p. 103–107). For the area of Bluff and Comb Ridge in southeastern Utah, Gregory (1938, p. 58, 59) has divided the Morrison formation from bottom to top into the Bluff sandstone member, the Recapture shale member, the Westwater Canyon sandstone member, and the Brushy Basin shale member. Stokes (1944, p. 959, 960, 962, 963) considers that the Salt Wash sandstone member correlates with the lower part of the Recapture shale member and that the Bluff sandstone member "is a local phase of the widespread Entrada formation." The Entrada age of the Bluff sandstone member has not been accepted by most geologists familiar with the beds in question. However, serious consideration may be given to the suggestion by Goldman and Spencer (1941, p. 1745, 1759, 1765) that the Bluff sandstone member is equivalent to the Junction Creek sandstone of southwestern Colorado and is a pre-Morrison type of sand stone, probably of Summerville age. These authors (1941, p. 1765, 1766) suggest that the greenish clay, or clay shales, and greenish clayey material common in the sandstones of the Morrison formation may be "altered fine volcanic ashes." This suggestion is supported by the local abundance of bentonite in the Morrison, its rarity in the upper part of the Curtis formation, and its absence in the older Jurassic beds of the Colorado Plateau area.

2. The Wanakah formation as now defined (Fischer, 1942, p. 368; Baker, Dane, and Reeside, 1947, p. 1664, 1668) includes a sequence of limestone, gypsum, sandstone, and shale between the Entrada sandstone and Morrison formation in northeastern Arizona, northern New Mexico, southwestern Colorado, and southeastern Colorado. Its base is defined by the Pony Express limestone member in southwestern Colorado and by the Todilto limestone member elsewhere (Silver, 1948, p. 77). In many places the basal limestone, or limestone and gypsum, is overlain by considerable sandstone and shale. Locally the limestone member is underlain by some gypsiferous sandstone, as at Cuchillo Blanco about 40 miles north-northwest of Albuquerque, New Mexico. The contact with the Entrada is sharp in some sections, as near Durango, Colorado, and gradational in others, as near Thoreau and Gallina, New Mexico. In eastern New Mexico the Todilto limestone member locally appears to grade both vertically and laterally into sandstones typical of the Entrada, as shown on the Canadian Escarpment southeast of Las Vegas, west of Gallegos, and on Alamogordo Creek near Tucumcari (C. B. Read and Ernest Dobrovolny, personal communication). The contact with the Morrison formation is more difficult to define except where the base of the Morrison is marked by a massive sandstone. In eastern New Mexico the Wanakah formation may include some 30 to 70 feet of sandy shale and sandstone bearing red and green chert concretions (Goldman and Spencer, 1941, p. 1762–1764; Dobrovolny and Summerson, 1946) at present assigned to the Morrison formation.

The Wanakah formation is considered to have been deposited under subnormal marine conditions at the southern margin of the sea of Curtis time (Heaton, 1939, p. 1161, 1176; Baker, Dane, and Reeside, 1947, p. 1668), as shown by the presence of fish of marine affinities, by the inclusion of thick deposits of gypsum, and by stratigraphic position nearly identical with that of the Curtis formation. Proof of its marine character is based mainly on the presence of fish remains at Bull Canyon, Guadulupe County, New Mexico (Koerner, 1930), on Felch Creek, 15 miles north of Canon City (Dunkle, 1942), Colorado, and on the Piedra River in southwestern Colorado (Dunkle, personal communication). Whether the Wanakah formation represents all of the Curtis formation where fully developed, as in the Uinta Mountains, or only part will be difficult to prove at most places. Where its upper part includes beds similar to the Summerville formation, as at Redrock, Arizona, and Beclabito Dome, New Mexico, it probably represents all of Curtis time. Throughout much of northern New Mexico its gradational relationship with the underlying Entrada sand-

stone and its generally sharp contact with the Morrison formation may indicate that only lower Curtis time is represented. However, the very presence of thick masses of gypsum in the Todilto and Pony Express limestone members of the Wanakah formation shows that there could have been little, if any, erosion before Morrison time. Perhaps the absence of gypsum in the Wanakah formation west of Laguna, New Mexico, is related to pre-Morrison erosion. Farther north in the area extending from Durango to Ouray, Colorado, the Wanakah formation includes beds at the top resembling the Summerville and probably represents all of Curtis time.

As now understood, the Wanakah formation includes a sequence of limestone, gypsum, sandstone, and shale extending at least from Tuba, Arizona, on the west to eastern New Mexico. In western Colorado it extends northward as far as Ouray and Ridgeway and may extend to Bostwick Park west of Vernal Mesa, where Reeside and Williams (Personal communication) noted about 30 feet of gypsum, thinly laminated limestone, shale, and sandstone below beds typical of the Morrison formation. The gypsiferous sequence from top to bottom includes 8 feet of soft, light-buff, fine-grained sandstone, 5 feet of massive white gypsum, 20 feet of pinkish sandy shale, and 2 to 10 feet of gypsum interbedded with thinly laminated, dark-gray limestone. Locally the basal 4 feet consists of yellow sandstone that is replaced laterally by gypsum.

In eastern Colorado the Wanakah formation extends northward along the Front Ranges to near Morrison where it has been included in the Ralson formation by LeRoy (1946, p. 47–57). The Ralston formation as defined includes a shale-marl facies and a gypsiferous facies. These facies occupy distinct areas and have not been traced into each other, judging by LeRoy's descriptions. Apparently they have been correlated with each other because they occur at the same stratigraphic position. Such a correlation may be valid but needs to be confirmed by more extended field work. Possibly the shale-marl facies is younger than the gypsiferous facies. This is suggested by the presence of *Unio*-like pelecypods from bed 12 of the type section of the Ralston formation at Ralston Reservoir 5 miles north of Golden (LeRoy 1946, p. 51). These fossils, collected by Fred Swain, were examined by Dr. Teng-Chien Yen, who compared them with *Vetulonaia faberi* Holt, a species from the Morrison formation of western and northwestern Colorado.

Marine connection of the Curtis sea with the marginal area of Wanakah deposition most likely occurred in southern and southeastern Utah, because there the outcrops of the Wanakah formation roughly parallel the adjoining southern limits of the Curtis and Summerville formations over a distance of 200 miles. Direct marine connection of the Wanakah formation of southwestern Colorado with the Curtis formation farther north near Snowmass Canyon, Wolcott, and State Bridge seems improbable because the two areas are separated by a positive mass on which the Morrison formation overlaps onto Precambrian rocks. A seaway in eastern Colorado is even less likely, because the northernmost outcrops of the Wanakah formation are nearly 100 miles south of the southern limit of the Curtis formation (Heaton, 1939, p. 1156, 1157; Reeside, 1931).

3. The term Entrada sandstone is employed for a widespread sandy facies in southern and eastern Utah, northern Arizona, northern New Mexico, the panhandle of Oklahoma, and much of western, northern, and eastern Colorado (Gilluly and Reeside, 1928, p. 76; Baker, Dane, and Reeside, 1936, p. 7, 8, 46, 49; Heaton, 1939, p. 1157–1176). The Entrada is absent over much of central and south-central Colorado and has not been identified in Kansas. The name has been used for red beds at a similar stratigraphic position in Wyoming and the western part of the Dakotas, but its suitability there is questionable except in southernmost Wyoming. Even excluding these States from consideration, it is second only to the Morrison formation as the most widespread Jurassic lithologic unit in the western interior region. In general, the Entrada sandstone has been readily and correctly identified except in northern New Mexico and southeastern Colorado, where the earlier geologic studies indicated it was absent. It is now known to be widely present in New Mexico (Baker, Dane, and Reeside, 1947, p. 1664–1668) in the upper part of the Wingate sandstone as defined by Dutton (1885, p. 136,

137) and to be represented in southeastern Colorado by the Exeter sandstone (Darton, 1928, p. 306; Heaton, 1939, p. 1165, 1166; Stovall, 1943, p. 51–57). In the Iron Springs district of southwestern Utah the Entrada sandstone includes the arkosic sandstone and maroon shale and siltstone at the base of the Pinto sandstone of Leith and Harder (1908, p. 37; Mackin, 1947, p. 6, 7).

Extension of the term Entrada sandstone north of Colorado and Utah should take into account the persistence of a similar sandstone at essentially the same stratigraphic position. To most geologists the Entrada sandstone connotes a widespread, light-colored sandstone facies characterized by clean, rounded quartz grains that are generally of two sizes, by conspicuous color banding of red, pink, gray, and white, and by large-scale tangential cross bedding. The larger grains are generally well rounded and frosted and are scattered through a matrix of much smaller and remarkably even-sized grains. This facies constitutes most of the Entrada sandstone throughout its extent (Baker, Dane, and Reeside, 1936, p. 7; Heaton, 1939, p. 1175; Thomas and Kruger, 1946, p. 1277). However, it passes westward rather rapidly in southwest Utah, in the San Rafael Swell of east-central Utah, and in the Uinta Mountains of northeastern Utah into a red-bed facies characterized by dark-red, silty sandstone that is well bedded, or plane-bedded, but not cross-bedded. The red-bed facies includes the type Entrada on the San Rafael Swell and is essentially identical lithologically with the Preuss sandstone of north-central and northeastern Utah, eastern Idaho, and westernmost Wyoming. The Preuss of Idaho and Wyoming possibly has more dull-red or reddish-gray units than the Preuss or silty Entrada of Utah, but they resemble each other much more closely than either resembles the light-colored, clean, sandy facies of the Entrada. Problems of nomenclature would have been much simplified if the early workers in the San Rafael Swell had given separate names to the two facies of the Entrada. Unfortunately they did not realize that the silty facies persisted far to the north and that the eastward change to the eolian facies occurred constantly within a distance of about 30 miles. In practice recent workers in northern Utah and Wyoming have consistently referred the deep-red silty facies to the Preuss and the light-colored, clean sandstone facies to the Entrada.

In Wyoming, therefore, a sandstone facies of the Entrada has been identified in the southernmost part of the State from the Baxter Basin on the west (Thomas and Kruger, 1946, p. 1279, 1281) to the Front Ranges on the east (Reeside, 1931, p. 1093–1103), but farther north a different terminology seems to be in order, as is discussed under annotation 18.

The age of the Entrada sandstone must be based partly on its stratigraphic position between the Carmel formation and the Curtis formation, whose ages are discussed under other annotations, and partly on the fact that it appears to be represented in Montana (Cobban, 1945, p. 1262, 1285–1290; Imlay, 1947, p. 258, 259) by a disconformity whose duration corresponds to the middle and upper Callovian. Of course, locally the lower part of the Entrada sandstone may be equivalent to the upper part of the Carmel formation elsewhere, or the upper part of the Entrada may be equivalent to Curtis elsewhere. Certainly the Moab tongue of the Entrada sandstone is of lower Curtis age (Baker, Dobbin, McKnight, and Reeside, 1927, p. 804). Since the source of the sand in the tongue appears to have been to the east or southeast (McKnight, 1940, p. 89, 97, 98), the Entrada sandstone of west-central Colorado is probably also partly of Curtis age. Very likely a similar kind of sand was derived from highlands in central Colorado, all during Entrada and Curtis time.

The character of the contact between the Entrada sandstone and Curtis formation has been variously interpreted. In most places the contact is sharp. Locally in east-central and in southwestern Utah (McKnight, 1940, p. 94–96; Baker, Dane, and Reeside, 1936, p. 8) and in the Uinta Mountains (Thomas and Kruger, 1946, p. 1279) considerable scouring of the upper surface of the Entrada sandstone is evident. However, such features may be expected wherever marine deposits transgress continental deposits and do not necessarily mean cessation of deposition or any appreciable time not represented by sediments. In northwestern Colorado in the area of Snowmass Canyon and State

Bridge the contact of the Entrada with the Curtis appears to be transitional, although it can be selected within a foot or two and is mappable. The Entrada in this area may actually be of lower Curtis age, corresponding in origin to the Moab sandstone tongue. In that case a transitional contact would be expected. Where the Entrada is overlain directly by the Morrison formation, such as in northwestern Colorado between Scofield Park and Meeker, or along the Front Ranges of Colorado north of Morrison, it seems likely that the top of the Entrada marks an appreciable disconformity. In areas where the Curtis formation is fully developed, such as in northern Utah, the disconformity at the top of the Entrada probably represents only a small part of the Callovian stage.

4. The Carmel formation has not been accurately dated, but it probably corresponds to most of the Middle Jurassic and the earliest Upper Jurassic. This assignment is indicated by some marine fossils from its lower part, by its stratigraphic position, and by its lateral gradation northward into the Twin Creek limestone, whose basal beds in southwestern Wyoming have furnished ammonites of Bajocian age and whose higher beds have furnished fossils of Callovian age. Several field workers have concluded that the Carmel formation grades laterally into the Twin Creek limestone, that the underlying Navajo sandstone is equivalent to the Nugget sandstone, and that the overlying Entrada sandstone is equivalent to the Preuss sandstone (Baker, Dane, and Reesid, 1936, p. 6, 7, 44–46, 54; Heaton, 1939, p. 1172–1176; Thomas and Kruger, 1946, p. 1275–1281; Bartram, 1930, p. 337–338, and 1940, p. 117–118). Intertonguing of the gypsiferous red beds, limestones, and sandstones of the Carmel with the normal marine shales and limestones of the Twin Creek implies that the Carmel was deposited in marginal areas of the Twin Creek sea. The problem is whether the Carmel is equivalent to all or only part of the Twin Creek limestone and whether the Carmel represents the same time interval throughout its extent. Thomas and Kruger (1946, p. 1281) would correlate the Carmel in the eastern part of the Uinta Mountains with the Gypsum Spring formation of Wyoming because both are "gypsiferous red bed sequences resting directly on the Navajo or Nugget sandstone . . . ." However, considering the gradational relationships of the Carmel with the Entrada sandstone at most places (Baker, Dane, and Reeside, 1936, p. 7) in Utah and the equivalence of the latter with the Preuss sandstone of the Wyoming-Idaho border area, it seems reasonably certain that the upper part of the Carmel formation is equivalent to the upper part of the Twin Creek limestone and consequently to part of the "Lower Sundance" of central Wyoming. Correlation of the lower part of the Carmel with the Gypsum Spring formation and the basal units of the Twin Creek limestone also seems probable, at least in those parts of Utah where the Carmel contains marine limestone near its base, as on the San Rafael Swell. The Homestake limestone of the Iron Springs district in southwestern Utah (Leith and Harder, 1908, p. 24–37) was originally referred to the Carboniferous but is now known to be identical with the Carmel (Mackin, 1947, p. 6).

The common fossils from the lower part of the Carmel formation have been listed by Reeside (*in* Gilluly and Reeside, 1928, p. 74, 75), who noted that the species occurred elsewhere "in the Ellis formation of Montana and in the lower part of the typical Twin Creek formation" (Baker, Dane, and Reeside, 1936, p. 7). This occurrence has been confirmed by Imlay, who found them in Montana, Wyoming, and Idaho associated with ammonites in beds ranging from upper Bajocian to lower Callovian age. One ammonite was obtained from the basal 6 to 12 feet of the Carmel on The Wedge along the San Rafael River, Utah. This was identified by Reeside as a macrocephalitid of Callovian age. The identification was substantiated by Imlay (1945, p. 1026), who noted that the ammonite resembled the macrocephalitid *Arctococeras* from the basal Callovian of Greenland and northern Russia. However, later collections by Reeside and Imlay from a limestone bed about 25 feet above the base of the Carmel formation at the head of Buckhorn Wash adjacent to The Wedge furnished some fragmentary ammonites unlike any known from the Upper Jurassic but resembling forms in the middle part of the Gypsum Spring formation near Cody (Imlay, 1947, p. 242). The faunal evidence obtained to date is not positive but rather strongly suggests

that the lower part of the Carmel is locally equivalent to the Gypsum Spring.

Bearing on this correlation is the observation of some geologists (Personal communication by A. A. Baker and John B. Reeside, Jr.) that the base of the limestones in the lower part of the Carmel formation represents a time plane marking a rapid transgression of the sea from the north and west and that the sea became more and more restricted subsequently during Carmel time. Similarly the Gypsum Spring formation represents a rapid, widespread transgression of the sea eastward from the main part of the Jurassic trough in eastern Idaho in early Middle Jurassic time, as shown by its wide distribution (Imlay, 1947, p. 236, 237) and by the presence of upper Bajocian fossils near the base of the Twin Creek limestone just above red beds that Love and Imlay consider equivalent to the type Gypsum Spring formation. Correlation of the beginnings of these transgressions seems logical.

5. The probable ages of the Navajo sandstone, Kayenta formation, and Wingate sandstone have been discussed in detail by Baker, Dane, and Reeside (1936, p. 5, 6, 56–58). The evidence afforded by the few reptilian remains is inconclusive (Camp, 1936, p. 52; Brady, 1935). If the basal part of the Carmel formation is of Bajocian age, as now seems likely, then the Navajo cannot be younger than Lower Jurassic, and the chances of the Kayenta and Wingate being Jurassic are considerably less than if the Carmel were entirely of Upper Jurassic age. Most geologists who have studied these formations agree that the Navajo sandstone resembles the overlying Jurassic sandstones lithologically, in distribution, and in westward thickening, whereas the Kayenta and Wingate in these respects resemble the underlying Chinle formation of Triassic age. However, until more detailed studies of these formations have been made in New Mexico and Arizona, it seems unnecessary to alter current age assignments.

6. The term Glen Canyon group undivided is employed by the U. S. Geological Survey for those parts of northeastern Arizona and northwestern New Mexico where the separate formations of the Glen Canyon group have not been recognized. It is the impression of several geologists that south of Red Rock, Arizona, the Navajo sandstone and Kayenta formation have disappeared, leaving only the Wingate sandstone (Baker, Dane, and Reeside, 1947, p. 1667). However, this conclusion implies a certainty not warranted by the amount of field work. Therefore, until detailed stratigraphic studies have been made it seems preferable to employ the noncommittal term Glen Canyon group undivided.

7. The Upper Jurassic age of the Curtis formation is based on a few small collections from east-central and southwestern Utah and fairly large collections from the eastern part of the Uinta Mountains in northeastern Utah and northwestern Colorado (Gilluly and Reeside, 1928, p. 79; Reeside, 1923, p. 38, 43, 44). From the basal part of the Curtis formation in the Uinta Mountains Reeside obtained ammonites belonging in the *Cardioceras cordiforme* zone of lower Oxfordian age. It seems apparent that the Curtis where overlain by the Summerville formation in east-central Utah represents only the lower part of the Curtis of northern Utah.

8. The Winsor formation was defined by Gregory (1948, p. 235–237) for a sequence of evenly bedded, friable, buff, white, and red sandstones overlying the marine Curtis formation along the valley of the Paria River, southern Utah. He notes that the Winsor formation occupies the stratigraphic position of the Morrison formation, but that lithologically it is quite different and that the Morrison formation disappears before reaching the Paria River. The red banding noted in the Winsor formation suggests correlation with the Summerville rather than the Morrison.

9. The age of the Summerville formation is based on its lateral gradation into the Curtis formation and complete replacement of the Curtis in parts of southeastern Utah and southwestern Colorado (Baker, Dane, and Reeside, 1936, p. 9; McKnight, 1940, p. 102). Rock types characteristic of the Summerville are present as far south as Beclabito dome in northwestern New Mexico and in near-by Red Rock, Arizona, but are there placed in the Wanakah formation because the Todilto limestone member is present. Goldman and Spencer (1941, p. 1759) have called attention to the resemblance between

the Summerville formation and the upper marl member of the Wanakah formation in the La Plata Mountains, southwestern Colorado. The Summerville is described as being darker, thinner-bedded, less calcareous, and finer-grained (Stokes, 1944, p. 986).

10. Beds similar to the Morrison formation, ranging from 1300 to 1800 feet or more in thickness, exposed along the northern and western margins of the Wasatch Plateau in central Utah were designated as Morrison (?) by Spieker (1946, p. 122–126). His query was based on a failure to find fossils, on the presence of rocks of similar characteristics in the overlying Indianola group of Upper Cretaceous age, and on the known thicknesses of the Morrison formation in eastern Utah (Baker, Dane, and Reeside, 1936, p. 51), which do not suggest a marked thickening toward central Utah. It would not be surprising if Spieker's Morrison (?) beds were of Lower Cretaceous age, because to the north along the Idaho-Wyoming border the Lower Cretaceous continental beds are thousands of feet thick (Mansfield, 1927, p. 101–108), and the Morrison is absent. Probably an intense search for macro- and microfossils in the Morrison (?) formation of Utah would assure a satisfactory separation from Lower Cretaceous beds of similar lithologic aspect, as it has in parts of Wyoming.

11. The 10,000 feet, or more, of Jurassic shale exposed along the northern and western margins of the Wasatch Plateau were originally included by Spieker (1946, p. 123–125; Eardley, 1933, p. 330–334) in the Arapien shale and were divided into the Twist Gulch member at the top and the Twelvemile Canyon member below. Subsequently the term Arapien shale was restricted by Hardy (1952, p. 91–95), with the approval of Spieker, to the rocks formerly included in the Twelvemile Canyon member. He likewise raised the term Twist Gulch member to formational rank. Lithologically most of the Twist Gulch formation is very similar to the Preuss sandstone of north-central Utah as exposed in the Central Wasatch Mountains and in the western part of the Uinta Mountains. However, the upper 175 feet in Salina Canyon (Hardy, 1952, p. 97) consists of fossiliferous shales and sandstones probably correlative with the Curtis formation in the writer's opinion. The restricted Arapien shale resembles both the Twin Creek limestone of northern Utah and the Carmel formation of eastern Utah but is more shaly than either. Its lower part has yielded a few fossils resembling species in the Carmel formation.

12. The Stump sandstone (Mansfield and Roundy, 1916, p. 81; Mansfield, 1927, p. 99, 100; Kirkham, 1924, p. 22) is lithologically similar to the Swift formation of western Montana, is essentially continuous with it through the Yellowstone National Park area, and contains a marine fauna including *Cardioceras*, of Oxfordian age. The term Stump was used by Thomas and Kruger (1946, p. 1278, 1279) for glauconitic beds in the western part of the Uinta Mountains as far east as Lake Fork. Farther east equivalent beds are much more calcareous and have been called Curtis in conformity with usage in Colorado.

13. The Preuss sandstone has been used for a thick sequence of red beds immediately underlying the Stump sandstone in eastern Idaho, westernmost Wyoming, and north-central Utah. Its exposures in the western part of the Uinta Mountains are lithologically similar to the type Entrada, but it passes eastward into a light-colored, clean sandstone facies to which most geologists apply the name Entrada.

14. The Twin Creek limestone (Veatch, 1907, p. 56, 57; Mansfield, 1927, p. 97, 98; Horberg, 1938, p. 11, 19, 20; Dobrovolny, 1941, p. 430–433) comprises seven distinct mappable members, as indicated on the chart, throughout its extent in western Wyoming, southeastern Idaho, and north-central Utah as far south as Thistle. The basal member is absent in the Uinta Mountains, but the upper six members persist from the area of Peoa eastward to Lake Fork. From Lake Fork eastward to near Vernal the three lower members of these six gradually wedge out, the fifth member from the base thins to several feet of limestone and gray shale, and the upper two members change laterally into red beds, which from White Rock eastward are commonly referred to the Carmel formation. The lower four members, which form about one-third of the Twin Creek limestone, are of Middle Jurassic age, and the upper three members are of lower Callovian age. These determinations are based on the presence of middle

Bajocian ammonites in the second member from the base, on Callovian pelecypods in the fifth and sixth members from the base, and on the lateral tracing of the members into areas in Montana and central Wyoming where fossil evidence has been obtained.

15. The term Nugget sandstone has been extended from southwestern Wyoming and eastern Idaho through southern Wyoming to the southern edge of the Black Hills in South Dakota. Because of this usage it seems preferable to extend the term to the Uinta Mountains and restrict the term Navajo to the probably equivalent, similar sandstones south of the Uinta Basin.

The Nugget sandstone cannot be younger than Lower Jurassic because it underlies beds of Bajocian age in eastern Idaho and western Wyoming. If it is of Lower Jurassic age the environment of deposition must have changed remarkably between northern Idaho and southernmost Alberta, where the lower 150 feet of the Fernie formation is characterized by phosphate and contains ammonites representing most of the Lower Jurassic and the basal part of the Middle Jurassic (Warren, 1934, p. 65, 66). However, the environment may not have been very different from that in western Nevada during part of Lower Jurassic time (Muller and Ferguson, 1939, p. 1618).

16. The Junction Creek sandstone was described by Goldman and Spencer (1941, p. 1750–1759, 1765–1767) for a clean, massive, whitish sandstone generally showing eolian cross bedding, locally attaining 500 feet in thickness, and nearly confined to the La Plata Mountains area of southwestern Colorado. They placed it in the Morrison formation, following current practice, but believed that it was older than the Morrison formation of east-central Utah and was probably equivalent to the Bluff sandstone member of the Morrison at Bluff in southeastern Utah. The Junction Creek sandstone is now considered distinct from the Morrison because its cleanness and large-scale cross-bedding resemble certain sandstones of Curtis or Entrada age much more than the impure clayey sandstone so characteristic of the Morrison formation throughout the western interior region.

17. It seems likely that the Stump sandstone is overlain directly by Lower Cretaceous strata in southeastern Idaho and westernmost Wyoming. The Ephraim conglomerate was once considered Jurassic (Imlay, 1945, p. 1020) on the basis of two small collections of marine fossils obtained by W. W. Rubey from limestones in conglomeratic beds that he then considered to represent the basal part of the Ephraim. Later work by Rubey (Personal communication) has shown that the fossils are actually from pebbly beds in the upper part of the Stump and that the Ephraim rests unconformably on the Stump or older beds.

18. The term "Lower Sundance" formation is being used provisionally by field geologists in central and southeastern Wyoming for a generally nonglauconitic sequence of oölitic limestone, calcareous gray shale, and fine-grained gray to red sandstones underlying a highly glauconitic sequence and overlying the Gypsum Spring formation or older beds (Love *et al.*, 1945). Toward the southeast and south the "Lower Sundance" becomes thinner, more sandy, and generally redder. Toward the northwest it becomes thicker, more shaly, more calcareous, and dominantly gray. Fossils from the lower part of the formation just above a persistent red unit on the north side of the Wind River Basin show the presence of the *Arcticoceras* and *Gowericeras* zones of the lower Callovian. The underlying red unit and still older limestone and gray shale at the base of the "Lower Sundance" are older than the type Sundance formation of the Black Hills area and have been traced westward into fossiliferous beds of Middle Jurassic age.

The sandstones in the upper part of the "Lower Sundance" are plane-bedded, very fine-grained, gray below and dull red above. The gray sandstone locally contains glauconite and limestone beds and appears to pass northward and westward into marine shale. It is correlated with the Hulett sandstone member of the Sundance formation in the Black Hills area (Imlay, 1947, p. 255). The red zone at the top persists southward and eastward a considerable distance but is absent at the northwest end of the Wind River Basin and throughout most of the Big Horn Mountains (Love *et al.*, 1945). Its disappearance northward has not been satisfactorily explained (Imlay, 1947, p. 258). It is

somewhat similar to the earthy red facies of the Entrada sandstone of Utah (Bartram, 1930, p. 335–346, 667–679; 1940, p. 117), to certain units of the Preuss sandstone of the Idaho-Wyoming border area, and to the Lak member of the Sundance formation in the Black Hills area. Compared with these units, its color is a lighter red and less persistent. It is quite different from the light-colored, clean sandstone facies of the Entrada sandstone in Colorado and eastern Utah. Extension of the term Preuss sandstone into central Wyoming seems inadvisable because, according to J. D. Love (Personal communication), there is a large area west of the Wind River Basin in which the "Lower Sundance" formation does not contain red beds in its upper part. Extension of the term Lak member of the Sundance formation into central and southeastern Wyoming is practicable.

Some gray shales underlying the gray sandstone are lithologically similar to the Stockade Beaver shale member of the Sundance formation and are characterized by an abundance of *Gryphaea nebrascensis* Meek and Hayden. Beneath them in the western part of the Wind River Basin occurs a thin unit of gray oölitic limestone interbedded with siltstone and shale that have furnished *Arcticoceras*, of lower Callovian age. These oölitic beds appear to pass eastward in the Big Horn Mountain area and the Hartville uplift into oölitic sandstones identical with the Canyon Springs sandstone member at the base of the Sundance formation in the Black Hills area (Imlay, 1947, p. 247). Below the oölitic beds occurs a red, gypsiferous unit and then a gray limestone and shale unit, which Love *et al.* (1945) place at the base of the "Lower Sundance" formation. The lowest limestone and shale unit has been traced westward by Love to the Jackson Hole area, where it has furnished *Defonticeras* and stephanoceratids of middle or upper Bajocian age.

19. The Middle Jurassic age of the Gypsum Spring formation has been well substantiated (Love *et al.*, 1945; Imlay, 1945, p. 1025; 1947, p. 241, 242). Recently J. D. Love and associates have traced the Gypsum Spring formation westward from the type locality at Red Creek at the west end of the Owl Creek Mountains to the Jackson Hole area south of Yellowstone National Park. Imlay has connected these occurrences with the basal brecciated red bed and limestone member of the Twin Creek limestone and has obtained *Defonticeras* and *Stemmatoceras* from limestones immediately above the basal red-bed member. As these ammonites represent the middle or upper Bajocian, the Gypsum Spring formation and its equivalents should represent some part of the middle or lower Bajocian. It probably does not represent any part of the Lower Jurassic judging by the various genera of pelecypods and gastropods obtained from it near Lander, Wyoming. (*See* list by Imlay *in* Love *et al.*, 1945, Mill Creek section.) As the Gypsum Spring formation represents the basal deposits of a transgressive sea, it may not be of the same age throughout its extent and may represent much more time at one locality than at another.

20. Imlay considers that the rocks assigned to the Gypsum Spring formation by U. S. Geological Survey parties in the Big Horn Basin of Wyoming are identical with the Piper formation of eastern Montana and that they include much more than the type Gypsum Spring formation as defined by Love (1939, p. 42, 45; 1945, sec. 4). This observation has been disputed by Love (Personal communication) for the sections along the west side of the Big Horn Mountains. Evidently the problem can be settled only by additional field work and field conferences.

21. The Piper formation includes all the Middle Jurassic beds underlying the Rierdon formation in eastern Montana east of the Sweetgrass-Big Belt-Bridger line of uplift (Imlay, 1948). It was at first identified as Gypsum Spring formation (Imlay, 1947, p. 236, 237, 242, 243), but later field work demonstrated that the Gypsum Spring formation of the type area (Love, 1939, p. 42, 45; 1945, sec. 4) in central Wyoming represents only the basal Middle Jurassic and correlates with the lower member of the Gypsum Spring formation, as employed by U. S. Geological Survey parties in Montana and in parts of the Big Horn Basin of Wyoming. This usage in Montana arose because it was practical in mapping and because the Gypsum Spring formation was assumed to include all the beds of Middle Jurassic age older than the type Sundance formation. As

the beds equivalent to the type Gypsum Spring in Montana are not mappable, the name Piper is employed in that State for beds hitherto called Gypsum Spring. The Piper formation is characterized by a lower red bed-gypsum member, middle gray shale-limestone-dolomite member, and an upper red bed-gypsum member. The middle member has furnished such ammonites as *Defonticeras* and *Teloceras*, which are of middle or upper Bajocian age. The upper red-bed member grades laterally in western and northern Montana into yellowish siltstone, sandstone, and limestone that contains *Arctocephalites* and *Procerites* and is considered to be of upper Bathonian age. These members grade vertically into each other, and to some extent the red beds grade laterally into the middle member.

22. The term "Upper Sundance" formation is being used provisionally in central and southeastern Wyoming for a highly glauconitic sequence of shale, calcareous sandstone, and limestone underlying the nonmarine Morrison formation (Love *et al.*, 1945). The lower part is mostly shale or sandy shale, and the upper part mostly sandstone or limestone. The sequence is somewhat more sandy than the equivalent Redwater shale member of the Sundance formation in the Black Hills area and much less sandy than the Stump sandstone of western Wyoming. An upper Oxfordian age for its lower part is shown by the presence of *Cardioceras cordiforme* (Meek and Hayden) and related species.

23. The term Sundance formation is still employed in north-central and northeastern Wyoming (Pierce and Andrews, 1941, p. 111–113), western Nebraska (Condra and Reed, 1943, p. 20–22), and the Dakotas (Ballard 1942, p. 1562) for dominantly marine Upper Jurassic beds underlying the Morrison formation and overlying the Gypsum Spring formation or, in places, the Nugget sandstone. It probably can be divided into two mappable units corresponding with the Rierdon and Swift formations of Montana. The various members noted in the Black Hills area (Imlay, 1947, p. 244–273) will be useful over a large part of eastern Wyoming. Correlation of these members faunally is based on the presence of *Cardioceras* in the lower part of the Redwater shale member and of *Arcticoceras* in the Canyon Springs sandstone member and the basal part of the Stockade Beaver shale member.

24. The Swift formation in Montana (Cobban, 1945, p. 1281–1286; Imlay, 1947, p. 260, 261) is correlated with the Redwater shale member of the Sundance formation in the Black Hills, the Curtis formation of Utah, the Stump sandstone of eastern Idaho and westernmost Wyoming, and the "Upper Sundance" formation of central Wyoming because of the widespread occurrence of species of *Cardioceras*, *Goliathiceras*, and *Pachycardioceras* in the lower parts of these formations. The Swift formation may actually include older beds than the other formations listed, as its basal beds in the Little Rocky Mountains and the Bearpaw Mountains have furnished *Bourkelamberticeras* and *Pavloviceras* beneath the zone of *Cardioceras cordiforme*.

25. The Rierdon formation of Montana (Cobban, 1945, p. 1277–1281; Imlay, 1947, p. 230, 231, 250, 252, 256) contains four ammonite zones of lower Callovian age characterized from bottom to top by (1) *Arcticoceras*, (2) *Gowericeras*, (3) a finely ribbed form of *Kepplerites* associated with *Cadoceras*, and (4) a coarsely ribbed form of *Kepplerites* associated with *Gulielmiceras* and *Gulielmites*. The lowermost zone characterized by *Arcticoceras* has been identified elsewhere at the base of the Sundance formation near Cody, Wyoming; in the basal sandstone of the "Lower Sundance" formation in the Hartville uplift of southeastern Wyoming; in the Canyon Springs sandstone member and basal part of the Stockade Beaver shale member of the Sundance formation in the Black Hills; in an oölitic limestone in the "Lower Sundance" formation on Red Creek near the west end of the Owl Creek Mountains of central Wyoming; and in a similar oölitic limestone overlying a red-bed unit in the Gros Ventre area south of Yellowstone Park. The widespread distribution of this genus at the base of a marine sequence directly above a red-bed unit at many places indicates a rapid spreading of the sea at the beginning of the Upper Jurassic.

26. The Middle Jurassic age of the Sawtooth formation is based on the presence of *Defonticeras* near its base and of *Arctocephalites* and *Procerites* in its upper silty beds, and on

its stratigraphic position beneath beds containing varied ammonite faunas that allow accurate correlation with the lower Callovian ammonite zones of northwest Europe. The formation pinches out by overlap on the South Arch of the Sweetgrass uplift and on the Big Belt uplift east of Helena but grades eastward into the Piper formation in northern and southern Montana.

27. The upper Portlandian age of the Viñales limestone has been firmly established by a detailed study (Imlay, 1942) of its fossils, which were obtained from many localities in Pinar del Rió, Santa Clara (now Las Villas), and Camagüey provinces in Cuba and from many stratigraphic levels ranging from a few feet to hundreds of feet above its base. Palmer (1945, p. 7–9) has divided the Viñales limestone on a lithologic basis into two formations—the Quemado formation below and the *Aptychus* beds or his restricted Viñales limestone above. He points out that most of the Portlandian ammonites described by Imlay are from the Quemado formation and most of the aptychi are from the overlying *Aptychus* beds. As the affinities of the aptychi are clearly with the Neocomian of Europe, Palmer considers that the age of the *Aptychus* beds is most likely Neocomian, although still open to question. However, in several collections studied by Imlay (1942, p. 1428) the ammonites and aptychi are associated and in some cases are even on the same rock slab. Also, some of the collections consisting mostly of aptychi have Portlandian ammonites such as *Pseudolissoceras* and *Virgatosphinctes*. It appears, therefore, that aptychi identical with Neocomian forms in Europe occur in Cuba in beds of latest Jurassic age. However, considering the great thickness of the Viñales limestone, the finding of Lower Cretaceous fossils in the upper part of the formation would not be surprising. The Viñales limestone is reliably reported to be overlain in one place in Pinar del Rio Province by beds containing *Orbitolina texana* (Roemer), which is indicative of the upper part of the Lower Cretaceous.

28. Ammonites of late Oxfordian age (Burckhardt, 1930, p. 61, 62; Spath, 1931, p. 400, 592, 593) have been obtained from concretions in shales or shaly limestones at the base of steep-sided limestone hills (mogotes) in the northern part of Pinar del Rió Province, Cuba. Dickerson and Butt (1935, p. 116–118) considered that the fossiliferous concretions were in the upper part of the Cayetano (or San Cayetano) formation. However, Palmer (1945, p. 6, 7) found microfossils of Upper Cretaceous age in the Cayetano formation and consequently postulated that the Oxfordian ammonites were actually in the basal part of a thrust sheet resting on the Cayetano formation. He applied the name Jagua formation to about 400 feet of very thin-bedded, shaly limestone from which the ammonites were obtained. As the Jagua formation is overlain directly by the Viñales limestone of upper Portlandian age, it seems evident that they are separated by an unconformity representing the Kimmeridgian and part of the Portlandian.

Imlay has recently examined numerous ammonites from the Jagua formation in the collections of Dr. Mario Sanchez Roig. These ammonites include *Phylloceras*, *Euaspidoceras*, *Ochetoceras*, *Vinalesphinctes*, and the subgenera of *Perisphinctes* belonging to *Arisphinctes*, *Dichotomosphinctes*, *Discosphinctes*, and *Biplices*. The species are very similar to forms in the *Peltoceras transversarium* and *P. bimammatum* zones of southern and central Europe. Most of the perisphinctid ammonites do not range above the upper Oxfordian, although *Discosphinctes* is common in the lower Kimmeridgian. If the lower Kimmeridgian was represented, there should be some specimens of *Idoceras*, or of *Sutneria* of the group of *S. platynota* Reinecke, or other genera such as are common in beds of that age in Mexico and southern Europe.

29. The Todos Santos beds (Sapper, 1937, p. 26–28) of northernmost Central America have been shown by Müllerried (1942a, p. 129; 1942b, p. 470–475) to contain Lower Cretaceous mollusks in their upper part and Jurassic plants in their lower part. The plants are reported to be identical with species occurring in the Lower and Middle Jurassic beds of western Oaxaca. Marine fossils of Jurassic age have not been found, and stratigraphic sections have not been measured. Farther south near Tegucigalpa, Honduras, the Upper Jurassic has been identified on the basis of a few ammonites similar to

Macrocephalites (Letter from Otto H. Hass dated March 24, 1944).

30. The late Upper Jurassic is represented in eastern Veracruz and adjoining areas by gray, brownish, or yellowish, platy, thin-bedded limestone and some black shale that have yielded fossils of Kimmeridgian and Portlandian ages (Burckhardt, 1930, p. 97, 98). Beds of upper Oxfordian age have not been identified but are possibly represented by the *Nerinea*-bearing limestone recorded by Müllerried (1942b, p. 475) from southeast of Cintalapa in western Chiapas.

31. Red beds with associated salt and gypsum that were penetrated in a well near Tonalapa in the Chinameca area west-northwest of Minatitlán in southeastern Veracruz (Huntley, 1923, p. 1165; Burckhardt, 1930, p. 96–98) are overlain concordantly by limestone and shale containing Kimmeridgian and Portlandian fossils. Similar red beds wherever observed in Tamaulipas, Veracruz, Oaxaca, and Puebla are overlain concordantly by Upper Jurassic marine limestone (Burckhardt, 1930, p. 95–97); Baker, 1930, p. 167–169; Burckhardt and Müllerried, 1936, p. 315, 316). The lower limit of the red beds has not been observed but is apparently not older than late Callovian considering that most of the Callovian is well represented by thick marine sediments. The distribution of the salt facies from Guatemala and Tabasco on the east to western Puebla and Oaxaca on the west is shown by the presence of many salt springs throughout that area (Sapper, 1937, p. 44, 45; Wittich, 1926, p. 115–128) and by salt domes in the Isthmus of Tehuantepec (Ver Wiebe, 1926, p. 348–358).

32. The Lower and Middle Jurassic rocks of Chiapas, Oaxaca, and southern Puebla belong mainly to a coal-bearing continental facies (Müllerried, 1933, p. 41, 42, 79; 1936, p. 36, 37; Buckhardt, 1930, p. 29; Wieland, 1914, p. 10–17), whose age has been determined mainly by means of plants, although its upper part in western Oaxaca has furnished ammonites of Middle Jurassic age.

33. In Cerro de la Virgin, south of Tlaxiaco in western Oaxaca, the basal Valanginian bituminous shale and limestone (Felix, 1891, p. 141; Burckhardt, 1930, p. 99) containing *Subthurmannia* and *Neocosmoceras* is underlain disconformably by about 300 feet of medium- to thick-bedded coralline limestone and that by at least 600 feet of thin-bedded to shaly gray limestone, which is considerably higher stratigraphically than the Oxfordian beds exposed on Cerro Titania west of Tlaxiaco.

34. A little north of Amoltepec in western Oaxaca occur black, thin-bedded limestone and carbonaceous shale that contain the ammonite *Idoceras* and are lithologically nearly identical with the *Idoceras*-bearing beds near Mazapil, Zacatecas (Flores, 1909, p. 115; Burckhardt, 1930, p. 99).

35. In Cerro Titania west of Tlaxiaco in western Oaxaca limestone, sandstone, and marl have furnished echinoderms, brachiopods, and pelecypods (Felix, 1891, p. 172–180, Pl. 27), of which some species are identical, according to Burckhardt (1930, p. 99), with species in the Oxfordian Zuloaga limestone near Mazapil, Zacatecas. These beds are underlain (Felix, 1891, p. 140) by varicolored, light-gray, greenish, or reddish marls and brown limonitic intercalations containing ammonites of Middle Jurassic or early Upper Jurassic age.

36. Marine shale, marl, limestone, and some sandstone of Callovian age have been found (Burckhardt, 1930, p. 26, 32, 35, 36, 43; Burckhardt and Müllerried, 1936, p. 310–312) in the gorge of the Rió Tecocoyunca north of Cualac, Guerrero, and near Tezoatlán and Mixtepec, Oaxaca. They have furnished ammonites representing both lower and upper Callovian (Burckhardt, 1927), but the exact stratigraphic positions of the various species have not been determined. Spath (1933, p. 769) considers that the highest beds characterized by *Erymnoceras mixtecorum* are not as high as the zone of *Peltoceras athleta*.

Elsewhere in Mexico marine Callovian has been identified only in the Huasteca area at the bottom of the gorge of the Rio Nexcotla several kilometers below the electric power plant. The evidence consists of large specimens of *Subgrossouvria* similar to *S. neogea* (Burckhardt). These were obtained from a dark gray, rather massive limestone that rests unconformably on an igneous intrusive rock and is overlain with probable disconformity by Upper Jurassic black claystone and thin-bedded limestone. The stratigraphic position of the Callovian lime-

stone with respect to the red Huizachal formation has not been ascertained because the latter does not outcrop in the gorge.

37. A section of coal-bearing beds near El Consuelo in the Tezoatlán region of western Oaxaca was measured by Wieland (1914, p. 10–17) and assigned to the Lower Jurassic. He recognized a lower division about 672 feet thick, consisting mainly of black shale but including some sandstone and coal, and an upper division about 1187 feet thick consisting of interbedded sandstone, conglomerate, shale, and coal. Burckhardt (1927, p. 92–94; 1930, p. 28, 32) has shown that the upper division contains ammonites of early Middle Jurassic age and is overlain by beds containing ammonites of late Middle and early Upper Jurassic age. Therefore, it is doubtful whether the Lower Jurassic includes more than the lower division of plant-bearing beds. Some of the plants from the lower division have been identified by Lozano (1916) with plants at the top of the marine Lower Jurassic sequence in the Huesteca region of Veracruz.

38. North of Cualac in northeastern Guerrero dark marl and marly limestone underlying beds of Callovian age have been identified by Burckhardt (1930, p. 26, 36) as Bathonian on the basis of the ammonites *Sirenoceras paracontrarium* (Burckhardt) and *Eurycephalites boesei* (Burckhardt). The underlying beds are sandstones bearing lenses of coal and many plants and oysters of undetermined age. They are underlain by conglomerate and quartzitic sandstone that rest on mica schist.

39. The latest Jurassic rocks exposed in the mountains bordering the Huasteca area consist of brown tuffaceous beds, yellow to gray bentonitic shale, lenses of gray to brownish-gray limestone, and a few thin beds of black chert. These beds were named the Pimienta formation by Arnold Heim (1926, p. 84–86; 1940, p. 332–334) after a small village south of Tamazunchale near the Río Moctezuma. The type Pimienta formation as exposed along the Mexico City-Laredo highway differs from that farther south near Huauchinango, Teziutlán, and Almanza by containing more limestone beds and less tuffaceous material. The top of the formation appears to be transitional into the Tamaulipas limestone within 10 to 30 feet. Heim supposed the Pimienta formation to be of Portlandian age, but he did not find fossils. Imlay subsequently found many ammonites of upper Portlandian age at many localities, including the type section. The most common ammonite genera are *Parodontoceras*, *Substeueroceras*, *Himalayites*, *Corongoceras*, *Hildoglochiceras*, *Pseudolissoceras*, and *Durangites*. The Pimienta formation is characterized by tuffaceous material and high bituminous content and is the source of many oil seeps.

40. The Tamán formation was proposed by Heim (1940, p. 334) for 1500 feet or more of black, thin- to medium-bedded limestone interbedded with dark shaly limestone and calcareous shale underlying the Pimienta formation along the Río Moctezuma southwest of Tamazunchale. Its Kimmeridgian age is shown by the presence of the ammonites *Haploceras fialar* (Oppel), *Sutneria*, and *Aspidoceras* and the pelecypod *Aulacomyella*. Similar black limestone and shale, probably 600 to 1000 feet thick, crop out in the mountains west of the Huasteca area between Tamazunchale and Huauchinango (Burckhardt, 1930, p. 91–94). This is underlain near Huayacocotla by about 400 feet of thick-bedded limestone of probable Oxfordian age that provisionally is included in the Tamán formation.

41. The Huayacocotla formation includes the 1300 feet, or more, of marine Lower Jurassic in the Huasteca area of northern Veracruz and Puebla (Imlay, Cepeda, Alvarez, and Diaz, 1948). It consists mainly of dark claystone and siltstone but includes interbeds of sandstone and conglomerate and a few lenses of limestone. It is overlain unconformably by red beds. Ammonites occur throughout most of the formation. Plants occurring in the basal beds in association with ammonites and at the top of the formation, where ammonites are lacking, furnish a correlation with the lower part of the continental facies of western Oaxaca. A generalized section of the formation was described by Burckhardt (1930, p. 12–20; Table 2). Descriptions of the ammonites are now awaiting publication. The name Potrero shale, originally proposed by Böse (1898, p. 175) for these marine Lower Jurassic beds, was overlooked until recently and was applied to a Lower Cretaceous formation in Sonora (King,

1939, p. 1660). The name Potrero shale was abandoned in favor of Huayacocotla formation, with the approval of the interested Mexican geologists.

42. Wells drilled near Tampico (Muir, 1936, p. 15, 16, figs. 8, 9; Burckhardt, 1930, p. 95, 96) penetrated from 450 to 600 feet of late Upper Jurassic sedimentary rocks consisting mainly of gray and brown limestone but including some black, carbonaceous, and bituminous shale and marked basally by a conglomerate that rests on red beds. Ammonites are common throughout and date the beds as middle Kimmeridgian to upper Portlandian. The pelecypod *Aulacomyella*, not known below the middle Kimmeridgian, occurs directly above the basal conglomerate. A disconformity may be present at the top of the Jurassic, as the *Substeueroceras* beds do not appear to be represented, and the basal beds of the Cretaceous contain considerable glauconite.

43. The La Caja formation (Imlay, 1938, p. 1659, 1683) consists of dominantly dark, thin-bedded limestone, marl, and shale, is the offshore equivalent of La Casita formation, and has an extensive distribution throughout the central part of the Mexican geosyncline. Its contacts with overlying and underlying formations are distinct but conformable, and the succession of ammonite faunas indicates that deposition was continuous in the central part of the Mexican geosyncline into Cretaceous time. According to the ammonite record (Burckhardt, 1930, p. 64, Tables 4–6; Imlay, 1939, p. 19–24), the La Caja formation represents the Kimmeridgian and Portlandian stages.

44. The Zuloaga limestone (Imlay, 1938, p. 1657, 1679) is the offshore equivalent of the La Gloria formation but was also formed near shore wherever bordering landmasses were low. It outcrops extensively throughout the central part of the geosyncline as well as in the large anticlinal uplifts of the Sierra Madre Oriental from the region of Saltillo southward. Its contact with the underlying red beds is generally marked by a conglomerate but in some sections is transitional. Its upper part is definitely of Argovian age, but its basal part may be as old as upper Divesian. The Novillo formation of Heim (1940, p. 319, 320) is lithologically very similar to the Zuloaga limestone, occupies the same stratigraphic position, and was not adequately defined until 1940.

45. The Huizachal formation includes the 1400 feet or less of red beds immediately underlying the normal marine Upper Jurassic in eastern and northern Mexico, overlying marine Lower Jurassic in the Huasteca area of Veracruz, Puebla, and Hidalgo, and overlying Paleozoic or probable Pre-cambrian rocks at many other places. Both upper and lower contacts of the Huizachal formation appear to be erosional unconformities. The Huizachal formation has not furnished fossils, but correlation with the red beds and salt of southern Mexico is indicated by lithologic characteristics, stratigraphic position, and enough recorded occurrences to show that the red beds of the two areas are parts of the same formation.

46. The La Casita formation (Imlay, 1936, p. 1110; 1937, p. 600–602; 1938, p. 1683) includes near-shore clastic, carbonaceous, and gypsiferous beds of late Upper Jurassic age above the La Gloria formation or the Zuloaga limestone. It has an extensive distribution around the northern and eastern margins of the Mexican geosyncline and grades basinward into the less clastic La Caja formation. Its upper and lower boundaries are sharply defined in most sections and locally mark erosionally disconformities.

47. The Olvido formation (Heim, 1940, p. 320, 321) consists of gypsum interbedded with some limestone, dolomite, and variegated shale and sandstone. It is widely distributed in the Sierra Madre Oriental (Burckhardt, 1930, p. 87; Böse, 1923, p. 130, 202, 203). Its lithology, stratigraphic position, and geographic relationship to Upper Jurassic landmasses are apparently identical with those of the Buckner formation of the southern United States. Its stratigraphic position shows that it is older than the beds characterized by *Idoceras balderum* (Oppel). Its lateral gradation locally into the Zuloaga limestone suggests it is of late Oxfordian age.

48. Upper Portlandian ammonites (Burckhardt, 1930, p. 266; Kellum, 1937, p. 68–92, Pl. 9, figs. 1–6) were found by Hegwein in the eastern part of the San Carlos Mountains of Tamaulipas. The age determination is based on the presence of *Kossmatia* and of uncoiled

ammonites similar to species in the Viñales limestone of Cuba. Until a thick sequence of Jurassic rocks was discovered in the southern United States, some geologists doubted whether Jurassic ammonites were actually obtained from the San Carlos Mountains.

49. The La Gloria formation (Imlay, 1936, p. 1105; 1938, p. 1677, 1679) includes near-shore deposits of early Upper Jurassic age lying above red beds and below the La Casita formation. It is characterized by sandstone and conglomerate, although some sections consist mainly of thick- to medium-bedded limestone. It has been identified only in eastern Durango, southern Coahuila, and northeastern Chihuahua but may have an extensive distribution around the northern margin of the Mexican geosyncline. Its boundaries are generally sharply demarcated and locally represent minor erosional disconformities. Its upper part contains ammonites of late Oxfordian age (Burckhardt, 1912, p. 203, 204, 209–213).

50. The presence of lower Lias in west-central and southern Sonora in the upper part of the Barranca formation has been discussed by Burckhardt (1930, p. 23, 24, 41, 42), King (1939, p. 1645–1659, Pl. 5), and Wilson and Rocha (1946, p. 23–29). They show that the Lias fossils occur in shale and limestone interbedded with considerable sandstone that is as much as 2500 feet thick and that these beds pass eastward and southeastward into about 1500 feet of gray, quartzitic sandstone that appears to be mostly nonmarine. King (1939, p. 1659) suggests that "the northeastern margin of the Barranca basin must have crossed northeastern Sonora at some point near or south of Nacozari, extending thence southeastward across western Chihuahua."

The extension of the Lias into northwestern Sonora is shown by the discovery of an *Arnioceras* by Donald White of the U. S. Geological Survey. The ammonite was obtained above a conglomeratic sequence about 4 kilometers southeast of La Linaña mine in the vicinity of El Antimonio, Sonora.

51. The Malone formation (Cragin, 1897, p. 817–819; Wilmarth, 1938, p. 1273) includes the Upper Jurassic rocks in Malone Mountain and in the hills north and east of Malone (Torcer) station in Hudspeth County, Texas. The formation consists of near-shore deposits (Albritton, 1938, p. 1758–1761) similar to the La Casita formation of northern Mexico but basally includes a thick mass of gypsum that occupies the same stratigraphic position as the gypsiferous Olvido formation of Mexico and the Buckner formation of the southern United States.

52. The Cotton Valley group, including the Bossier and Shuler formations (Swain, 1944, p. 579), is comparable lithologically and stratigraphically with the La Casita formation of northern Mexico (Imlay, 1940a; 1940b; 1943). The Bossier formation (Swain, 1944, p. 582) has furnished ammonites of lower Kimmeridgian age near its base and of middle Kimmeridgian age near its middle (Imlay, 1941; 1945; Swain, 1944, p. 592, 593, 610, 611). The Shuler formation (Swain, 1944, p. 594) has furnished Jurassic pelecypods and on the basis of stratigraphic position is considered of Portlandian age (Imlay, 1943, p. 1472; Swain, 1944, p. 599).

53. The Buckner, Smackover, and Eagle-Mills formations are closely related in depositional history. The Buckner formation is characterized by anhydrite and red beds and appears to be a back-reef facies of late Smackover time (Swain, 1949, p. 1207, 1219–1223). The Smackover formation consists of two facies. One facies contains considerable sandstone and shale interbedded with limestone, has been found only in northeastern Louisiana, and is similar to the La Gloria formation of northern Mexico. The other facies consists mainly of interbedded limestone and dolomite, is widespread from Alabama to Texas, and is similar to the Zuloaga limestone of northern Mexico. The upper Oxfordian age of the Smackover formation is based (1) on the presence of ammonites similar to species of *Dichotomosphinctes* and *Discosphinctes* from beds of late Oxfordian age in Mexico, (2) on comparisons of some of the pelecypods and gastropods with European species, and (3) on the presence of Kimmeridgian ammonites in the overlying Cotton Valley group.

The age of the Eagle Mills formation has been a subject of considerable debate as the formation has not furnished fossils. When the formation was first penetrated by drilling, it was compared with the Permian of west Texas

because of the presence of thick masses of salt (Spooner, 1932, p. 605). Later it was placed in the Upper Jurassic (Hazzard, 1939, p. 174, 175) because of the presence of similar beds in the Upper Jurassic of southern Mexico. Imlay (1940a, p. 12; 1943, p. 1436–1438) considered that it was probably lower Oxfordian, as indicated by the transition of the salt beds into the overlying Smackover limestone and by the presence of upper Oxfordian ammonites only 300 feet above the salt. Recently Hazzard, Spooner, and Blanpied (1947, p. 483–503) have summarized the information pertaining to the beds below the Smackover formation. They present evidence that most of the red beds, including the type section of the Eagle Mills formation, underlie the main sequence of salt and anhydrite instead of grading laterally into it, as previously considered. They show that some higher red beds and associated conglomerates overlying the salt mass transgress all older rocks onto Paleozoic rocks and are invariably associated with the basal part of the Smackover formation. They propose to call these 150 feet or less of gravelly red beds the Norphlet formation, the underlying 600 to 1300 feet of salt the Louann salt (equals Eagle Mills salt of previous papers), and the underlying 100 to 200 feet of anhydrite, red beds, and gravel the Werner formation (equals Louann tongue of previous papers). They infer from the stratigraphic and structural evidence presented that the red-bed facies of the Eagle Mills formation underlies the Werner formation unconformably and is older than the late Paleozoic Morehouse shale of northeastern Louisiana. They conclude that the entire evaporite sequence to the top of the Louann salt is more likely Permian than Jurassic. The question of the age of the Eagle Mills formation and associated beds needs further study in view of the remarkable resemblance of the Eagle Mills to the Huizachal formation of Mexico. Also, it is not evident why a Paleozoic age for the red-bed facies of the Eagle Mills would prove that the salt masses are Paleozoic. Proponents of a Permian age for the salt masses have not been able to explain how the salt escaped erosion during the long interval from Permian to Upper Jurassic (Swain, 1949, p. 1207).

54. The Portlandian age of the Knoxville formation has been amply confirmed (Stanton, 1895; Pavlow, 1907; Spath, 1923, p. 306; Crickmay, 1932; Burckhardt, 1930, p. 64; Anderson, 1945) and a correlation established with the Mexican faunal zones characterized by *Durangites* and *Substeueroceras*. The lowest Portlandian has not been identified faunally but may be present, as Anderson (1945, p. 940) mentions that the lower 1200 feet of the Knoxville has not furnished fossils other than *Aucella*. It is evident that the disconformity between the Knoxville and the Horsetown formations represents only a brief interval, as all the stages of the Upper Jurassic and Lower Cretaceous are represented. It probably was developed during the Infra-Valanginian (Berriasian), as that time has not been definitely identified in California. Anderson (1945, p. 924–932) considers that the Knoxville is a series and is divisible into three groups, which from lowest to highest he names the Newville, Grindstone, and Elder Creek. The justification for these groups is not apparent from his brief discussion of their lithologic features and faunal content and must await confirmation by field parties doing areal mapping.

55. The Franciscan formation has been discussed in detail by Taliaferro (1942; 1943), who concludes that it is a volcanic facies of late Jurassic age, that it grades upward into the dark shales and limestones of the Knoxville formation, and that in places its upper beds are equivalent to the lower part of the Knoxville elsewhere. He states that the volcanic rocks and radiolarian cherts typical of the Franciscan occur locally in the Knoxville and that Knoxville fossils have been found in beds that field geologists have mapped as Franciscan. These conclusions are at variance with the opinions of many earlier workers but are supported by a wealth of data and are based on many years of field work. They should be given preference until refuted by argument based on equally thorough investigations. Recently some geologists have reported the finding of an ammonite of Albian age in the Franciscan formation of the San Francisco area, but the discovery has not been verified in publication. To have much weight in evaluating the age of the Franciscan, it must be proven that the ammonite was found in place.

In the San Francisco area the Franciscan has been considered a group and has been divided into a number of formations (Lawson, 1914) that from base to top include the Cahil sandstone, Sausalito chert, Marin sandstone, Ingleside chert, and Bonita sandstone. The Cahil sandstone contains the Calera limestone member. These units have not been listed in the chart because of lack of space and because their existence outside the San Francisco area has not been demonstrated.

56. Intrusion of the granodioritic batholiths of the Sierra Nevada can be dated only as younger than the Mariposa slate of Oxfordian or early Kimmeridgian age and older than the Upper Cretaceous beds, which in places rest directly on granodiorite (Hinds, 1934, p. 182–192; Taliaferro, 1942, p. 104, 105). A late Jurassic to early Cretaceous age is favored because it probably took considerable time to erode the sedimentary rocks that were above the batholiths at the time of their intrusion. Many intrusive bodies have been given local names and assigned questionably to the late Jurassic (Jenkins, 1943, p. 673–676).

57. The Mariposa slate along the western side of the Sierra Nevada has been dated as lower Kimmeridgian on the basis of the occurrence of the ammonite *Amoeboceras* (*Amoebites*) and of finely striate Aucellas (Hyatt, 1894, p. 402; Smith, 1894; Reeside, 1919, p. 10). Taliaferro (1933, p. 149) raised the Mariposa to the rank of a group and divided it into the Mariposa slate above and the Indian Gulch agglomerates below. However, the name Mariposa should not be used for a group and also for a unit within that group. The upper part of the Mariposa slate in the Colfax quadrangle was separated by Smith (1910, charts opposite p. 217, 221) as the Colfax formation, apparently because it contained the ammonite *Perisphinctes colfaxi* Hyatt, which he and Crickmay (1931, p. 68) considered Portlandian. However, such an age assignment needs checking by additional fossil collecting, as the type of *P. colfaxi* is reported to be lost, and the original description and illustrations are not adequate for generic determination.

The Mariposa slate of the Colfax quadrangle, in the northern part of the Sierra Nevada, may include the equivalents of the Amador group (discussed below), because "*Olcostephanus*" *lindgreni* Hyatt, from half a mile south of Colfax, Placer County, greatly resembles finely ribbed forms of *Gowericeras*, such as occur in the lower Callovian in Montana. This suggestion supports the observations of Crickmay (1933a, p. 57).

Below the Mariposa slate Taliaferro (1933, p. 149) recognizes thousands of feet of Jurassic beds that he first called the Tuolumne group but later redefined and assigned to the Amador group (Taliaferro, 1942, p. 89, 90; 1943, p. 282–284). This group had previously been included partly in the Mariposa, partly in the Paleozoic Calaveras group, and partly in intrusive rocks of unknown age. According to Taliaferro (1943, p. 283) the Amador group is known to extend from the Middle Fork of the American River to Fresno County. Near the northern end of its extent on the Cosumnes and Calaveras Rivers it is 7100 feet thick and consists of the Cosumnes formation below and the Logtown Ridge agglomerates at the top. The Cosumnes formation rests unconformably on the Paleozoic. Near the southern end of the extent of the Amador group it is at least 14,000 feet thick, but its base is not exposed. This thickness on the Merced River has been divided into five formations including (1) lower volcanics, (2) pillow basalts, (3) Hunter Valley cherts, (4) Peñon Blanco volcanics, and (5) the Aqua Fria slates. Taliaferro correlates the Logtown Ridge agglomerates with the Peñon Blanco agglomerates and the Hunter Valley cherts with part of the Cosumnes formation. He notes that in most places the Amador group grades into the Mariposa slate, but that locally they are separated by an angular unconformity.

The Amador group is probably entirely of Upper Jurassic age. One perisphinctid ammonite was obtained from the Cosumnes formation about 2000 to 2500 feet west of its upper contact with the Logtown Ridge agglomerates on the right bank of the Cosumnes River about two miles below Huse Bridge, El Dorado County. Although perisphinctids existed as early as middle Bajocian, this specimen shows a backward curvature of its secondary ribs similar to that in *Grossouvria* of Callovian age.

Another perisphinctid ammonite was obtained from the Logtown Ridge agglomerates from a quarry on Logtown Ridge near the Huse

Bridge about 6 miles northwest of Plymouth and 35 miles east of Sacramento. This ammonite is similar to *Divisosphinctes* from the upper Oxfordian and to some of the finer-ribber species of *Pachysphinctes* from the Kimmeridgian. Its features exclude the perisphinctids of the Callovian.

Considering that the Mariposa slate contains ammonites of lower Kimmeridgian age, an Oxfordian age for the underlying Logtown Ridge seems reasonable.

58. The Monte de Oro formation (Turner, 1896, p. 548, 549) has been correlated with the Bajocian of Europe and with some beds in Douglas County, Oregon, on the basis of its flora (Ward, 1900, p. 340–377; Fontaine, 1905, p. 48–145; Knowlton, 1910). Its stratigraphic position relative to the other Jurassic formations of the Sierra Nevada cannot be determined, as it is an isolated occurrence. However, the probably equivalent beds in Douglas County, Oregon, could perhaps be placed stratigraphically by careful mapping. According to Taliaferro (1942, p. 93), they appear to be older than the Dothan formation.

59. Thick Jurassic sequences, supposedly older than the Mariposa slate and outcropping near the north end of the Sierra Nevada, have been called the Milton formation in the Downieville quadrangle and the Sailor Canyon formation in the Colfax quadrangle (Turner, 1894, p. 232–234). Taliaferro (1942, p. 99, 100), following Bruce Clark (Unpublished manuscript), considers these formations equivalent, records the presence of fossils of Upper Triassic to Middle Jurassic age in the lower 9500 feet of the Milton formation, and notes the resemblance of the Milton to the Jurassic section in the Taylorsville area about 23 miles north of the most northerly Milton exposures. As the upper 13,000 feet of the Milton formation has not furnished fossils, it may be partly of Upper Jurassic age.

60. The Jurassic section on Mount Jura in the Taylorsville area at the northern end of the Sierra Nevada was described by Diller (1892, p. 369–394; 1908a, p. 34–60), and the ages of the various formations were determined by Hyatt (1894, p. 400–412). The section was later studied by Crickmay (1933b), who proposed a number of new formation names and made several age assignments different from those made by Hyatt. Taliaferro (*in* Reed, 1943, p. 106) has expressed the opinion that the stratigraphic interpretation by Diller is essentially correct. For purposes of comparison, the sequence as determined by Crickmay is shown on the chart alongside the sequence as determined by Diller and Hyatt. The latter sequence has been modified slightly on the basis of age determinations by Muller and Ferguson (1939, p. 1612, 1613, 1622). The faunal evidence is sufficient to correlate the Hardgrave sandstone with the upper part of the lower Lias, the Mormon sandstone with most of the upper Bajocian, and the Bicknell sandstone with the lower Callovian. The Foreman formation is generally considered equivalent to the Mariposa slate (Taliaferro, 1942, p. 100), although Crickmay (1933b, p. 902, 914, Pl. 34) reports ammonites from it that are of Callovian age if correctly identified.

61. In the Redding district (Diller, 1906, p. 5) the Jurassic sequence consists of the Modin formation at the base and the Potem formation at the top. These are locally separated by the Bagley andesite. The Potem contains fossils that Stanton considered identical with species in the Hardgrave sandstone of the Taylorsville area. He considered the Modin formation basal Jurassic rather than Triassic, although the fossil evidence is not conclusive. Muller and Ferguson (1939, p. 1612, 1622, Table 3) have indicated similar tentative correlations. The term "Bend formation" was applied to the Jurassic sequence in the Lassen quadrangle (Diller, 1895, p. 4) but was later abandoned, because of prior usage, in favor of Potem formation.

62. Although the presence of Lower Jurassic rocks in western Nevada has been known for many years (Gabb, 1870, p. 5–18; Hyatt, 1892, p. 411; 1894, p. 418), information concerning their lithologic features and faunal content is due almost entirely to Muller and Ferguson (1939, p. 1609–1622). Description of the faunas of the Dunlap and Sunrise formations should help considerably in determining the Lower Jurassic sequences in other structurally complex areas of Nevada and California.

63. The Mesozoic sequence in southwestern Oregon was originally mapped by Diller (1903;

1907, p. 401–421; 1908b, p. 367–402), who correlated the Galice formation with the Mariposa, the Dothan formation with the Franciscan, and the unconformably overlying Myrtle formation with the Knoxville and Horsetown. He recognized that the Dothan was beneath the Galice but considered that the beds were overturned. Louderback (1905) recognized volcanics and chert similar to the Franciscan formation in the lower part of the Myrtle formation and separated that part as the Dillard series. Diller (1907) insisted that the lower part of the Myrtle contained the same fossils as the Knoxville formation of California, which was then considered Lower Cretaceous and definitely younger than the Franciscan. Subsequently, the Knoxville was proven to be of latest Jurassic age (Pavlov, 1907; Spath, 1923, p. 306; Crickmay, 1932, p. 1–11; Anderson, 1932, p. 320; 1945, p. 938–945), and the Franciscan was interpreted as a volcanic facies of late Jurassic age at least partly equivalent to the Knoxville (Taliaferro, 1942, p. 83–87; 1943) and not older than the Portlandian. This interpretation explains the finding of Knoxville fossils in beds resembling the Franciscan, such as in the Dillard formation in southwestern Oregon. Furthermore, if the Dillard is equivalent to the Franciscan, then the Dothan cannot be equivalent to the Franciscan. With these possibilities in mind, Taliaferro (1942, p. 81–83, 87–90, 107) re-examined the formations in question in Oregon and concluded that the Dothan underlies the Galice normally and is identical lithologically and stratigraphically with the Amador group exposed along the western side of the Sierra Nevada.

64. The Jurassic of east-central Oregon has been divided into a number of formations representing a considerable part of the system (Packard, 1928; Lupher, 1941). Reference of the Lonesome formation to the Upper Jurassic is tentative, as it has not furnished fossils. The Trowbridge shale is referred to the Callovian because of the presence of the macrocepalitid ammonite *Lilloettia* and of cosmoceratid ammonites similar to *Gowericeras*. The beds containing *Lilloettia* are probably of middle Callovian age, as the genus occurs in beds of that age in Alaska and has not been found in the lower Callovian beds of the western interior of the United States. The Izee group belongs in the middle or early upper Bajocian, as shown by the association of stephanoceratids and sonniniid ammonites. The genera listed by Lupher (1941, p. 259–263) suggest the same assemblage as occurs in the middle siltstone member of the Tuxedni formation in Alaska, in the Rock Creek member of the Fernie formation in Canada, and near the base of the Twin Creek limestone in western Wyoming. The Colpitts group has been assigned to the lower Bajocian because of the presence of abundant sonniniid ammonites and certain other ammonites and because of the absence of stephanoceratids, which are not known below the middle Bajocian (Lupher, 1941, p. 253–255). The Mowich group contains ammonites of late middle Liassic to early upper Liassic age (Lupher, 1941, p. 245–247). The upper part of the Donovan formation contains ammonites representing the early middle and late lower Lias.

65. The Shuksan formation (Weaver, 1945, p. 1392) includes some Upper Jurassic sandstones, shales, conglomerates, and cherty beds outcropping in the valley of the North Fork of the Shuksan River north of Mount Baker in north-central Washington. Evidence for an Upper Jurassic age comparable with that of the Mariposa slate consists of finely striated forms of *Aucella* that Stanton identified with *A. erringtoni* Gabb (Smith and Calkins, 1904 p. 27). Other Jurassic rocks may exist in Washington (Weaver, 1945, p. 1379, 1401), but faunal evidence of such is lacking.

66. The basal part of the Naknek formation is of lower Oxfordian age (*mariae* to *plicatilis* zones of northwest Europe), as shown by the presence of various species of *Cardioceras* in the Cook Inlet area, in the Matanuska Valley, and in the Alaska Peninsula. The overlying beds in the lower part of the Naknek formation are of probable upper Oxfordian age, as shown by the presence of *Amoeboceras* (*Prionodoceras*) associated with finely striate *Aucella bronni* (Rouiller). The subgenus *Prionodoceras* in Europe ranges from the *Perisphinctes cautisnigrae* zone to the *Rasenia cymodoce* zone but is essentially characteristic of the upper Oxfordian. Striate species of *Aucella*, such as occur with *Prionodoceras* in Alaska, are characteristic of the late Oxfordian and the Kimmerdgian. In

Russia they have been recorded in the zone of *Amoeboceras alternans*, which corresponds basally with the zone of *Decipia decipiens*. In Alaska, *Aucella* has not been found in association with *Cardioceras* except at one locality in the Puale Bay area. Assuming that the fossils at Puale Bay were collected from the same bed, they should represent the approximate boundaries of the lower and upper Oxfordian. A Kimmeridgian age to basal Portlandian age for the upper part of the Naknek formation is shown by the presence of *Aucella mosquensis* (von Buch) and *A. rugosa* (Fischer) in several thousand of feet of beds above the highest occurrence of *Amoeboceras* and *Cardioceras* (Martin, 1926, p. 175–180, 211–218, 231–232; Moffit, 1927, p. 38–42; Capps, 1927, p. 21–33).

The Naknek formation on the northern part of the Alaska Peninsula is overlain by Lower Cretaceous rocks, but near the southwestern end of the peninsula at Herendeen Bay, Port Moller, and Pavlof Bay it is overlain by the Staniukovich shale which is of Portlandian age on the basis of the occurrence of numerous *Aucellas* similar to *A. piochii* Gabb. These *Aucellas* have elongate, rather pinched umbones and are unknown in the Infra-Valanginian (Berriasian) of the Arctic region. The age significance of these *Aucellas* was not realized until after the correlation chart was drafted and ready for publication.

67. The Shelikof formation consists of a lower siltstone member 1500 feet or less in thickness, a middle sandstone member at least several thousand feet thick, and an upper siltstone member 700 to 1000 feet thick. The middle member contains some siltstone and conglomerate and grades laterally and vertically into the other members (Capps, 1922, p. 90–92, 97–102; Smith and Baker, 1924, p. 169, 176–178; Martin, 1926, p. 199, 200; Kellum, Daviess, and Swinney, 1945, p. 8–11, Figs. 6–9). At Wide Bay the lower member has furnished *Cadoceras*, *Kepplerites*, *Pseudocadoceras*, and *Lilloettia*. The last two genera were found only in the upper part of the lower member. At Puale Bay (formerly Cold Bay) the middle and upper members have furnished *Cadoceras* and *Pseudocadoceras*. The occurrence of *Pseudocadoceras* near the top of the formation on the mountain southwest of the head of Puale Bay (U.S.G.S. Mes. loc. 12387) shows that no part of the formation is younger than Callovian. The ammonites in the Shelikof formation are specifically identical with those in the Chinitna siltstone in the Cook Inlet area. The presence of *Reineckeia* (*Kellawaysites*) about 350 feet above the base of the Shelikof formation at one locality on Wide Bay shows that the basal part of the formation is not older than Callovian.

68. There has been considerable uncertainty whether the Kialagvik formation of the Wide Bay area (Capps, 1922, p. 91, 94–97; Smith and Baker, 1924, p. 169, 173–176) represents the latest Lower Jurassic or the lowest Middle Jurassic. (*See* discussion in Martin, 1926, p. 187–190.) Most paleontologists have favored a Lower Jurassic age, but Stanton (*in* Capps, 1922, p. 96, 97) considered a Middle Jurassic age more likely and thought that some of the collections indicated a correlation with the lower part of the Tuxedni formation of the Cook Inlet area. Recent collections from the Kialagvik formation show that it is equivalent to the lower part of the Tuxedni formation of the Tuxedni Bay area and is definitely of lower Bajocian age.

In the Kialagvik formation at Wide Bay three faunal assemblages can be recognized. The lowest assemblage is characterized by *Erycites kialagvikensis* (White), *E. howelli* (White), *Pseudolioceras whiteavesi* (White), and *Tmetoceras*. It contains a few representatives of *Hammatoceras* and *Sonninia*. The latter occurs at the top of the local range of *Tmetoceras* and the other species listed. This occurrence of *Sonninia* is probably significant, as the genus in Europe is not known below the top of the *murchinsonae* zone. The other genera in Europe occur in the Toarcian as well as in the lower Bajocian. The species of *Tmetoceras* match very closely with species in the lower Bajocian of Europe.

The next higher faunal assemblage in the Kialagvik is characterized by an abundance of the earliest stephanoceratid ammonite *Emileia* associated with *Sonninia*, *Erycites*, *Pseudolioceras*, and an oppeliid. Now *Emileia* in Europe is common in the *sowerbyi* zone and rare in the *sauzei* zone. Its association on Wide Bay with *Erycites* and *Pseudolioceras* suggests a position at the base of the *sowerbyi* zone. The

abundance of *Sonninia* and the presence of an oppeliid indicates an age not older than the *sowerbyi* zone.

The next higher faunal assemblage in the Kialagvik formation contains the ammonites *Stemmatoceras*, *Lissoceras*, *Leptosphinctes*, *Emileia*, and an oppeliid. The presence of the first three genera indicates an age not older than the *sauzei* zone. The same assemblage is well developed in the Cook Inlet area, where faunal and stratigraphic evidence indicates that it represents the lower part of the *sauzei* zone. This occurrence of *Leptosphinctes* is probably older than that at Sinai, Egypt, and considerably older than its occurrence in Europe at the base of the zone of *Parkinsonia parkinsoni*.

69. Lower Jurassic rocks at least 2300 feet thick have been identified on the northeast side of Puale Bay and at Alinchak Bay (Martin, 1921, p. 60; Capps, 1922, p. 90–94; Smith and Baker, 1924, p. 169–173; W. R. Smith, 1925, p. 195–196; Martin, 1926, p. 182–186; Kellum, Daviess, and Swinney, 1945, p. 3, 4, fig. 2). The Sinemurian is represented in several collections by *Coroniceras* and *Arnioceras*. Fragmentary specimens in various lots suggest the presence of the Hettangian, Pleinsbachian, and Toarcian stages.

70. The Chinitna siltstone ranges in thickness from about 2200 to 3500 feet and has furnished many determinative fossils (Martin and Katz, 1912a, p. 65–68; 1912b, p. 22; Moffit, 1922a, p. 142, 144; 1922b, p. 123, 124; 1927, p. 26–30; Paige and Knopf, 1907, p. 22; Chapin, 1918, p. 21, 33, 34). For purposes of locating the fossils stratigraphically it may be divided into three parts. Its upper 600 to 1100 feet consists of dark-gray rather massive siltstone that locally contains many sandstone lenses, is characterized by hard, thin calcareous beds that weather to a conspicuous yellow, is marked basally by a massive gray sandstone at Chinitna Bay and Tuxedni Bay, and contains the ammonites *Pseudocadoceras*, *Cadoceras*, and rare *Cosmoceras*. Its middle 1000 to 1600 feet consists of concretionary dark-gray siltstone that weathers reddish brown and has such characteristic ammonites as *Cadoceras*, *Pseudocadoceras*, *Paracadoceras* (rare), *Kepplerites*, *Lilloettia*, *Oxycerites*, and *Kheraiceras*. Its lower 200 to 1400 feet consists mostly of concretionary massive, rather soft siltstone, has some interbeds of fine-grained, greenish-gray sandstone, and contains such ammonites as *Kepplerites*, *Cadoceras*, *Paracadoceras*, *Lilloettia*, *Xenocephalites*, *Gowericeras*, and *Cosmoceras* (*Gulielmiceras*). Kirschner and Minard (1949), named this lower part the Tonnie member of the Tuxedni formation, but mapping has shown that the massive sandstone capping Tonnie Peak of the Iniskin Peninsula disappears to the south and that in most places the siltstones assigned to the Tonnie cannot be separated from the overlying siltstones of the Chinitna. Likewise, it was once considered feasible to map the upper 600 to 1100 feet of the Chinitna siltstone as a separate member from the underlying siltstone on the basis of a massive gray sandstone about 30 feet thick on Chinitna Bay and 80 feet thick on Tuxedni Bay. However, such a sandstone is absent over large areas, and no criteria for separating the siltstones could be established by field parties.

The lower part (roughly one-third) of the Chinitna formation is of lower Callovian age equivalent to the European zones of *Proplanulites koenigi* and *Sigaloceras calloviense*. This is indicated by an abundance of *Cadoceras* (*Paracadoceras*), which includes such European forms as *C. elatmae* (Nikitin) and *C. breve* Blake, by the absence of *Pseudocadoceras*, and by the rare occurrence of *Gowericeras* and *Cosmoceras* (*Gulielmiceras*). There is no evidence in southwestern Alaska for the existence of the *Arctococeras* zone of the basal Callovian or the still older *Arctocephalites* zone of the upper Bathonian. These are well developed in the western interior of the United States and Canada, in the Richardson Mountains in Yukon Territory, in northeastern Alaska and in East Greenland. Failure to find them in southwestern Alaska suggests an unconformity between the Chinitna siltstone and the Tuxedni formation.

The middle part (roughly middle third) of the Chinitna formation is considered middle Callovian as shown long ago by Pompeckj (1900). In Europe *Pseudocadoceras* ranges from the *calloviense* to the *coronatum* zone. *Cadoceras* ranges into the *coronatum* zone, and its earliest forms are from the upper Cornbrash of England. *Kepplerites* ranges from the *macrocephalus* to

the *calloviense* zone in England and apparently into the *coronatum* zone in northern Russia. *Lilloettia* is unknown outside of North America. *Paracadoceras* is represented in the middle part of the Chinitna formation by a single occurrence that is low stratigraphically. Most of the species of *Cadoceras* and *Kepplerites* in the middle part of the Chinitna are unlike those in the lower Callovian beds of Montana and are believed to be younger.

The upper part (roughly 600 to 1100 feet) of the Chinitna formation is correlated with the European zone of *Erymoceras coronatum* because of the presence of *Pseudocadoceras* and of compressed *Cadoceras* similar to *C. milaschevici* (Nikitin) and *C. stenolobum* (Keyserling) and by the absence of the other ammonites listed from the middle part. Furthermore, one specimen of *Cosmoceras* found as float near the base of the upper third of the Chinitna formation is similar to *C. spinosum* (J. de C. Sowerby) which is not known below the *Erymnoceras coronatum* zone and is more common at higher levels. The presence of *Pseudocadoceras* in the upper 100 feet of the formation on Iniskin Bay shows that no part of the formation is as young as the Oxfordian.

71. The Tuxedni formation includes 5000 feet or more of clastic sediments along the west side of Cook Inlet and in the Matanuska Valley (Martin and Katz, 1912a, p. 59–64; Eichwald, 1871, p. 138–200; Hyatt, 1896, p. 907, 908; Stanton and Martin, 1905, p. 397–402; Moffit, 1922a, p. 143, 144; 1922b, p. 120–122; 1927, p. 12–22; Mather, 1925, p. 165, 166; Martin, 1926, p. 140–158, 224–226, 278, 279; Paige and Knopf, 1907, p. 10, 16–19; Chapin, 1918, p. 21, 32, 33). In the Iniskin Peninsula on the outcrop it is divisible into six members (Kirschner, 1949), which from base to top include a basal siltstone and sandstone member, the Gaikema sandstone and conglomerate member, a siltstone member, the Cynthia Falls sandstone member, and the Bowser member. The two lower members contain ammonites identical with those in the uppermost part of the Kialagvik formation of the Wide Bay area. Older beds correlative with the lower part of the Kialagvik formation are present in the subsurface, as shown by the presence of an ammonite similar to immature forms of *Erycites kialagvikensis* (White) in cores at depths of 6339–6359 feet in the Iniskin Bay Association well near the head of Fitz Creek. The siltstone member just below the Cynthia Falls sandstone member is characterized by an abundance of *Normannites*, *Teloceras*, and *Chondroceras* (*Defonticeras*). With these are a few specimens of *Stephanoceras*, *Stemmatoceras*, and *Lissoceras*. This assemblage probably represents the upper part of the *sauzei* zone because *Teloceras* attained its greatest abundance in the upper part of that zone and because *Normannites* has not been recorded outside that zone.

The Cynthia Falls sandstone member of the Tuxedni formation has furnished a few fragmentary stephanoceratids. The overlying Bowser member is represented by two ammonite assemblages. The lower third or fourth of the member is characterized by *Sphaeroceras*, *Oxycerites*, *Stephanoceras*, *Leptosphinctes*, and *Lissoceras*. This assemblage represents either latest Bajocian or earliest Bathonian. In Europe *Leptosphinctes* and *Sphaeroceras* are typical of the *parkinsoni* zone of the latest Bajocian. *Oxycerites* ranges from the lowermost Bathonian into the Callovian. *Stephanoceras* (broad sense) ranges from the *sowerbyi* zone into the basal Bathonian. However, the forms from the basal part of the Bowser are very similar to *Ammonites linguiliferus* D'Orbigny, which Mascke has made the type of *Polyplecites* and which occurs in both the *parkinsoni* zone and the basal Bathonian. The presence of *Leptosphinctes* and *Sphaeroceras* favors assignment to the *parkinsoni* zone.

The upper two-thirds of the Bowser member is characterized by fairly abundant *Cranocephalites*, which differs from the slightly younger *Arctocephalites* by its last whorl becoming scaphitoid and remaining strongly ribbed. In east Greenland and other parts of the Arctic, *Cranocephalites* consistently occurs below *Arctocephalites*, but its exact age has not been determined because no fossils have been found in the immediately underlying beds. Fortunately, in southwestern Alaska well-preserved specimens of *Cranocephalites* have been found associated with *Reineckeia* and *Planisphinctes* immediately above the assemblage containing *Sphaeroceras*, *Leptosphinctes*, and *Stephanoceras*. The associations and stratigraphic relationships

show that *Cranocephalites* existed early in the Bathonian.

In the Tuxedni Bay area more of the Tuxedni formation is exposed than in the Iniskin Peninsula. The presence about 350 feet above the Lower Jurassic (?) volcanics of the ammonites *Tmetoceras*, *Erycites*, and *Pseudolioceras* shows that these beds are correlative with the lower part of the Kialagvik formation at Wide Bay. About 1000 feet higher a sequence of sandstone and siltstone, comprising units 22 to 36 of the section measured by Stanton and Martin (1926, p. 142, 143), contains abundant *Stemmatoceras*, *Sonninia*, and *Stephanoceras* and a few specimens of *Emileia*, *Defonticeras*, and *Lissoceras*. These probably represent the lower part of the *sauzei* zone. The overlying units 7 to 20 contain abundant *Teloceras*, *Normanites*, and *Defonticeras* that represent the upper part of the *sauzei* zone. Units 1 to 6 at Fossil Point on Tuxedni Bay contain *Defonticeras* and *Stephanoceras*, consist mostly of sandstone, and are probably correlative with the Cynthia Falls sandstone member. The overlying Bowser member, well exposed near the head of Bear Creek, contains the same faunal assemblages as on Iniskin Peninsula.

72. The Lower Jurassic is represented by tuffaceous beds on the Kenai Peninsula, where one collection (U.S.G.S. Mes. loc. 2981) contains an ammonite comparable with *Coroniceras*, of Sinemurian age. Similar tuffaceous beds on the west side of Cook Inlet have been referred to the Lower Jurassic because they appear to underlie the basal Middle Jurassic rocks conformably (Martin, 1926, p. 134–139; Moffit, 1927, p. 11, 12).

73. The Tordrillo formation, of the Alaska Range, has furnished several specimens of *Inoceramus* that appear to be identical with species in the lower part of the Tuxedni formation (Martin, 1926, p. 235). Capps (1935, p. 54) considered it impossible to separate generally the Todrillo formation from the underlying Skwenta group.

74. The Skwenta group, of the Alaska Range, has been assigned to the Lower Jurassic because of its stratigraphic position below the Tordrillo formation and because it is lithologically similar to the Talkeetna formation of the Matanuska Valley (Spurr, 1900, p. 149–152; Brooks, 1911, p. 85–87; Martin, 1926, p. 233, 234; Capps, 1926, p. 84–86; 1935, p. 59).

75. The Talkeetna formation, of the Matanuska Valley and the east side of the Talkeetna Mountains (Paige and Knopf, 1907, p. 16–19; Knolwton, 1916, p. 451–460; Chapin, 1918, p. 21, 29–32; Martin, 1926, p. 219–223; Capps, 1940, p. 72–74; 1927, p. 19–23), contains ammonites (U.S.G.S. Mes. loc. 6697 and 8569) that are provisionally referred to *Schlotheimia*, *Deroceras*, and *Xipheroceras*. The last two genera belong to the family Deroceratidae, which ranges through the Pliensbachian and Toarcian. The species present resemble those in the Pliensbachian.

76. Beds of upper Oxfordian or lower Kimmeridgian age have been identified in the Nutzotin Mountains on the basis of finely striate species of *Aucella* similar to *A. bronni* Rouillier (Moffit and Knopf, 1910, p. 16, 27–32; Martin, 1926, p. 242–245; Moffit, 1938a, p. 63–65; 1943, p. 125–127). Similar species of *Aucella* have been found a little farther north in the Slana-Tok district at the east end of the Alaska Range (Moffit, 1938b, p. 32). In the Chitina Valley district (Moffit, 1938a, p. 65, 66), beds of the same age are represented by rocks with *Amoeboceras* (*Prionodoceras*) associated with the finely striated *Aucella bronni* (Rouiller) (U.S.G.S. Mes. loc. 14495 and 14497). *A. rugosa* (Fischer) from the Nutzotin Mountains (U. S. Mes. loc. 16085) indicates a middle Kimmeridgian to basal Portlandian age. Some *Aucellas* from Mes. locs. 8111 and 18349 resemble *A. piochii* Gabb and are probably of Portlandian age.

77. The presence of Middle Jurassic rocks near the mouth of the Chitina River is implied by a few small collections of ammonites and pelecypods (Moffit and Knopf, 1910, p. 16; Martin, 1926, p. 239, 240). The fossils are too poorly preserved to allow positive identification, although they are certainly not older than the Jurassic.

78. Some 500 feet, or more, of tuffaceous conglomerate near Taral in the Chitina Valley has been identified as Lower Jurassic on the basis of stratigraphic position and lithologic characteristics, but it might be part of the Middle Jurassic sequence (Moffit, 1914, p. 17,

18, 25–27; 1938a, p. 64; Martin, 1926, p. 238, 239).

79. The latest Jurassic has been identified in southeastern Alaska only on Admiralty Island is a pebbly shale outcropping between two arms of Pybus Bay (Martin, 1926, pp. 377, 378). This pebbly shale has furnished many specimens of *Aucella* (Mes. locs. 3309, 8851, 10095, 10169) that belong to *A. piochii* Gabb or to similar species characterized by an elongate form and narrow umbones. Such *Aucellas* in California and in the Arctic region of Eurasia occur in beds of Portlandian age.

Somewhat older Aucellas occur on Kupreanof Island and Gravina Island. These were once assigned to the Lower Cretaceous (Martin, 1926, pp. 253, 379–382), but they include *A. rugosa* (Fischer) and *A.* cf. *A. mosquensis* (von Buch) which in Russia range through the upper Kimmeridgian into the basal Portlandian.

There is no paleontological proof for the existence of Jurassic rocks older than the Kimmeridgian in southeastern Alaska. However, a thick sequence lying on the Upper Triassic rocks should be of Jurassic age on the basis of stratigraphic position (Martin, 1926, p. 245–262; P. S. Smith, 1939, p. 40; Buddington and Chapin, 1929, p. 156, 157, 168–173).

80. Upper Jurassic beds have been identified faunally (Mes. loc. 20714) in the Kuskokwim region at only one place, which is 28.3 miles S. 8° W. of Aniak and 15.3 miles N. 42° E. of Nyac in the eastern part of the Akiak district. The fossils occur in a sequence of interbedded graywacke and green tuff and consist of well-preserved molds that the writer identifies with *Aucella mosquensis* (von Buch) and *A. rugosa* (Fischer). Some of the specimens of *A. mosquensis* have an articulating groove in the hinge area of the left valve, so there is no possibility of confusion with the similar *Aucellina* that occurs in beds of Albian age elsewhere in Alaska.

81. Lower Jurassic has been identified in the Barrow area of northern Alaska in at least six wells. It occurs in the South Barrow test well no. 3 at depths of 1645 to 2610 feet, in the Simpson test well no. 1 at depths of 5550? to 6300 feet, and in the Avak test well no. 1 at 1350 to 2310 feet. In the South Barrow test no. 3 the age determination is based mainly on ammonites obtained from cores. The species are very similar to species in the late Lower Jurassic of Europe, and the succession of ammonite genera is identical. Thus the beds from 1772 to 2063 feet correspond to the lower Toarcian, as shown by the presence of *Dactylioceras* and *Peronoceras*. The beds from 2069 to 2198 feet correspond to the upper Pliensbachian, as shown by the abundance of *Amaltheus*. In the Simpson test well no. 1 an ammonite fragment resembling an immature *Amaltheus* occurs in a core at 5680 feet, and another ammonite similar to *Metaderoceras* occurs at 5685 feet. The pelecypod *Oxytoma*, common in Jurassic rocks and rare in latest Triassic rocks, occurs at depths of 6174 to 6186 feet. Foraminifera from cores in both wells have been studied by Helen Tappan Loeblich who notes (Personal correspondence) that the assemblage is quite different from that in the overlying Lower Cretaceous rocks and resembles closely the Lower Jurassic Foraminifera of Europe. In the Avak test well no. 1 the ammonite *Coroniceras* has been obtained in a core at the depth of 1836 feet. The species is similar to *"Ammonites" bucklandi* Sowerby as figured by Wright (1878, Pl. 1, figs. 1–3). The presence of *Coroniceras* is strong evidence that the Sinemurian stage is represented. The similarity to *C. bucklandi* suggests correlation with the lower Sinemurian.

Middle Jurassic has been identified in the Topagoruk test well no. 1. One ammonite from a depth of 8111 feet belongs to *Pseudolioceras*. Another from a depth of 8113 feet probably belongs to *Tmetoceras*. The specimens are fragmentary but match very well with well-preserved specimens of these genera from the basal beds of the Kialagvik formation on Wide Bay in the Alaska Peninsula. Beds of about the same age are represented in the South Barrow test well no. 2 at depths of 2391 feet by a fragment that is probably an immature form of *Tmetoceras*. This genus occurs nearly world-wide in the highest beds of the Lower Jurassic and the basal beds of the Middle Jurassic and is particularly characteristic of the zone of *Ludwigia murchisonae*.

Upper Jurassic beds, representing the Upper Oxfordian to Kimmeridgian stages, have been identified by Foraminifera in the Topagoruk

test well no. 1 at depths of 6910 to 7820 feet. The age determination is based on identity of species with those in the highest Jurassic rocks that crop out in the foothills north of the Brooks Range.

The 5000 feet, or more, of the Kingak shale cropping out in northern Alaska was first identified as Jurassic on the basis of some fossils in its lower 1500 feet (Leffingwell, 1919, p. 103, 119, 120). These consisted mainly of *Erycites*, *Pseudolioceras*, and *Inoceramus* identical with forms in the Kialagvik formation (Martin, 1926, p. 262–265). Crinoid arms, obtained from the very base of the Kingak shale immediately above Upper Triassic beds, were compared by Springer (1926) with a European species of Lower Jurassic age. The crinoid determination has been confirmed by A. A. Bowsher, of the U. S. National Museum. Collections made north of the Brooks Range since 1945 contain ammonites from several levels in the Kingak shale and show that much of the Jurassic is represented.

The highest and most widespread part of the Kingak shale contains *Aucella rugosa* (Fischer), *A. mosquensis* (von Buch), and locally *A. bronni* (Rouiller). This part is mostly middle and upper Kimmeridgian but may also represent basal Portlandian judging by the ranges of those species of *Aucella* in the Arctic region.

The next lower part of the Kingak shale as exposed on the Canning River contains *Amoeboceras* (*Prionodoceras*) and the flattened, finely striate *Aucella bronni* (Rouiller) (U. S. Mes. loc. 22598, 21028). These represent either upper Oxfordian or lower Kimmeridgian. The subgenus *Prionodoceras* in Europe is the common form of *Amoeboceras* in the zones of *Perisphinctes cautisnigrae* to *Ringsteadia pseudocordata*.

Somewhat lower in the section on the Canning River have been obtained many specimens of *Cosmoceras castor* (Reinecke) (Mes. loc. 22597), which species in northwest Europe occurs in the zone of *Erymnoceras coronatum*. About 550 feet lower have been found fragments of *Arcticoceras* (Mes. loc. 22596) similar to *A. kochi* Spath. This genus occurs in the Western Interior of North America at the very base of the Callovian. An intermediate position in the Callovian is represented by *Pseudocadoceras grewingki* (Pompeckj), obtained west of the Canning River on the West Fork of the Ivishik River (Mes. loc. 22745). This species in the Cook Inlet area, Alaska, ranges through the middle portion of the Chinitna formation, which portion the writer correlates with the upper part of the *Sigaloceras calloviense* zone and the entire *Cosmoceras jason* zone of Europe.

The basal part of the Middle Jurassic has been identified by the ammonite *Pseudolioceras* on the Canning River and by *Pseudolioceras* and *Erycites* on the Sadlerochit River. On the Canning River these fossils were obtained probably 2000 feet lower stratigraphically than the collections of *Arcticoceras*.

Ammonites of Lower Jurassic age have been found near the base of the Kingak about 2 miles southeast of Lupine River. They consist of an *Amaltheus* similar to *A. engelhardti* (d'Orbigny) and a *Lytoceras* similar to *L. fimbriatum* Sowerby.

There is no faunal evidence for the upper Bajocian, the Bathonian, the latest Callovian, the lower Oxfordian, and the upper Portlandian. Those are probably represented by disconformities because mapping has demonstrated that the Callovian sequence laps across the early Bajocian and the Lower Jurassic and is in turn overlapped by the upper Oxfordian-Kimmeridgian sequence. The highest beds are overlain disconformably by the early Cretaceous Okpikruak formation.

82. The Upper Jurassic sequence in Milne Land has been described mainly by Rosenkrantz (1929, p. 135–154) and Aldinger (1935, p. 48–116), and the fossils have been determined by Spath (1935; 1936). Summaries of these studies, as well as others dealing with east Greenland, have been published by Koch (1935) and Teichert (1939). Sections of the Jurassic measured by Parat and Drach (1933) and Parat (1936) do not agree with those measured by Rosenkrantz and Aldinger and were probably based on much less field work.

The Upper Jurassic sequence probably includes basally the 200 meters of Charcot Bay sandstone (Spath, 1935, p. 63, 64, 67; Aldinger, 1935, p. 53–57), which has not furnished determinable fossils but grades into the overlying dark, sandy shales that contain *Cardioceras* aff. *C. zenaidae* Ilovaisky, of upper Oxfordian

age (Spath, 1935, p. 64, 65). Spath (1935, p. 66) places the beds of upper Oxfordian-lower Kimmeridgian age in the Kløft I formation, although the type section of that formation is on Kuhn Island far to the north. He notes that there is no natural boundary between it and the next overlying Cape Leslie formation. The Kløft I formation, as employed by Spath for Milne Land, includes from base to top about 200 meters of dark sandy shales, 70 meters of the *Pecten* sandstone, and about 240 to 250 meters of dark, sandy shale and marl. The lower shale unit has furnished *Cardioceras* and near its top some specimens of *Ringsteadia* and *Amoeboceras* (*Prionodoceras*). The basal part of the *Pecten* sandstone has likewise furnished *Amoeboceras* (*Prionodoceras*) (Spath, 1935, p. 64, 65). These are all considered upper Oxfordian. The lower 70 to 80 meters of the overlying shale has furnished *Amoeboceras* (*Amoebites*), *Rasenia*, *Pictonia*, and other fossils of lower Kimmeridgian age. Above follows 50 meters of unfossiliferous shale below a bed of crushed perisphinctids that Spath provisionally places at the base of the upper Kimmeridgian.

The Cape Leslie formation (Rosenkrantz, 1929, p. 147) as measured by Aldinger (1935, p. 51, 61–65) includes from base to top about 36 meters of unfossiliferous shale, 45 meters of reddish sandy marl with *Pectinatites*, 45 meters of dark micaceous sandy clays containing *Pallasiceras*, 25 to 50 meters of glauconitic sandstone and sandy shale containing *Pallasiceras* in the lower part and *Epipallasiceras* and *Behemoth* in the upper part, and 20 to 30 meters of sandy shale containing *Crendonites*. Spath (1936, p. 162) draws the boundary between the upper Kimmeridgian and the Portlandian within the glauconitic beds rather than at their top, mainly because of the presence of the ammonite *Behemoth*.

The Hartzfjaeld sandstone (Rosenkrantz, 1929, p. 149; Aldinger, 1935, p. 65–70) about 50 to 80 meters above its base contains the ammonites *Craspedites* and *Titanites?*, which Spath (1936, p. 162–164) considers of Portlandian age. At Hartz Mountain these are overlain within 8 meters by beds containing *Subcraspedites* resembling Lower Cretaceous species. About 95 meters higher at the same locality occurs *Craspedites* of late Jurassic aspect, which leads Spath (1936, p. 84–86, 163, 164) to suggest that the Hartzfjaeld sandstone may be partly lowest Cretaceous.

83. The probable occurrence of upper Kimmeridgian or Portlandian beds in Jameson Land is suggested by the presence of ammonites referred questionably to *Pectinatites?* and *Dorsoplanites* and perhaps by the presence of *Aucella mosquensis* (Von Buch) (Spath, 1936, p. 175–177).

84. For Jameson Land, Rosenkrantz (1929, p. 147) applied the term Koch Fjaeld formation to about 100 meters of shale and sandstone without determinable fossils overlying the Fossil Mountain formation, which he considered stratigraphically higher than the Vardekløft formation because of the presence of narrow-ventered ammonites that he assigned to *Quenstedtoceras*. Spath (1932, p. 124, 157; 1935, p. 70) has shown that these ammonites actually belong to *Arcticoceras* and occur in the lower part of the Vardekløft formation and that the so-called Fossil Mountain formation is a facies of the Vardekløft. This leaves the Koch Fjaeld formation as a possible representative of the upper Oxfordian and lower Kimmeridgian beds. However, Spath (1935, p. 9, 69) apparently prefers the term Kløft I formation for these beds, although Aldinger (1935) does not mention the term.

Regardless of stratigraphic terminology, the presence of beds of upper Oxfordian and lower Kimmeridgian age in Jameson Land has been proven faunally. *Cardioceras caelatum* Pavlow, from western Jameson Land, is correlated by Spath (1935, p. 68, 69, Pl. 15, fig. 3) with the zone of *Cardioceras cordatum*. Many impressions of *Amoeboceras* (*Prionodoceras*) have been found in southwestern Jameson Land (Spath, 1935, p. 25, 69, 74, 78, Pl. 13, figs. 5–7). Some *Amoeboceras* in southeastern Jameson Land (Aldinger, 1935, p. 36, 37) are considered by Spath (1932, p. 126; 1935, p. 69) to be closer to lower Kimmeridgian than to upper Oxfordian species of the genus.

85. The Vardekløft formation includes about 200 meters or more of black to gray, micaceous shales that include some finely sandy beds and one or more layers of calcareous concretions (Rosenkrantz, 1929, p. 145, 146; Aldinger, 1935, p. 22–26, 35, 36). Madsen (1904) con-

sidered the formation of Callovian age, but Spath (1932, p. 138–146) presented evidence that both the upper Bathonian and lower Callovian are represented. Spath's conclusions are confirmed by recent investigations of the Jurassic faunas of Alaska and the western interior of the United States, as discussed under annotations 25, 26, and 89. A stratigraphic break probably exists between the Vardekløft formation and the underlying Neills Cliff formation, of Lower Jurassic age, as Bajocian and lower Bathonian fossils have not been found in East Greenland. However, Spath (1932, p. 133, 138) points out that in northern Jameson Land the *Cranocephalites* beds of the upper Bathonian are separated from the top of the Lower Jurassic by more than 900 feet of unfossiliferous beds.

86. The Neills Cliff formation includes about 200 meters of sandstone, sandy shale, conglomerate, and impure limestone in the eastern part of Jameson Land and the southern part of Liverpool Land. It contains fossils of medial and late Lower Jurassic age (Rosenkrantz, 1929, p. 144, 145; 1934; 1942; Haug, 1926; Aldinger, 1935, p. 15–22). The lower 5 to 10 meters has furnished *Uptonia jamesoni* (Sowerby) of lower Pliensbachian age. Directly above at one locality *Beaniceras* sp. and *Lytoceras fimbriatum* (Sowerby) were obtained, which forms Rosenkrantz (1934, p. 48, 114) refers to the northwest European zone of *Tragophylloceras ibex*. Most of the formation is not very fossiliferous, but the upper 30 meters has furnished *Dactylioceras*, *Pseudolioceras*, and *Catacoeloceras*, of upper Toarcian age. Rosenkrantz (1929, p. 144, 145; 1934, p. 29) notes that the faunas characterizing the lower part of the formation were not found in the northern part of the area observed by him. This accords with the work of Arne Noe-Nygaard (1934, p. 63) in northeastern Jameson Land. This investigator found that the marine Lias, even at the base, contains mollusks that Rosenkrantz refers to the Toarcian.

87. The Cape Stewart formation occurs in eastern Jameson Land and near Cape Hope in southernmost Liverpool Land (Rosenkrantz, 1929, p. 143; Koch, 1929b, p. 251, 252; Hartz, 1896; Harris, 1926; 1931; 1932a; 1932b; 1935) and consists of 90 to 200 meters of nonmarine sandstone and shale that includes locally some limestone and conglomerate. It contains two distinct plant assemblages of which the lower has been assigned to the latest Triassic, and the upper to the basal Jurassic. Comparable deposits are reported from Trail Island and Geographical Society Island (Maync, 1947, p. 154).

88. Probably Trail Island and Geographical Society Island have a Jurassic sequence similar to that on Jameson Land, a short distance to the south, but few fossils have been definitely identified. The occurrence of the Bathonian in the southern part of Trail Island is shown by by presence of *Cranocephalites* in sandstones (Frebold and Noe-Nygaard, 1938, p. 25, 31). The Oxfordian or lower Kimmeridgian is represented on both islands by black shales containing *Amoeboceras* (Maync, 1947, p. 154, 155). The Portlandian is possibly represented in the northern part of Trail Island, according to Spath (*in* Frebold and Noe-Nygaard, 1938, p. 17).

89. On the west side of Kuhn Island, near Haakons Hytta, the Portlandian is represented by coarse sandstone and conglomerate containing species of *Aucella* and the ammonites *Pavlovia* and *Laugeites*, which have been identified by Spath (Maync, 1940, p. 24, 1947, p. 30, 136; Vischer, 1943, p. 16, 40, 41, 78, pl. 2). These beds, and probably equivalent beds in near-by areas, range from 75 to 160 meters in thickness, are referred by Maync (1947, p. 133–137) to the Rigi series, and are indicated to rest unconformably on older Jurassic rocks (1947, p. 184, 185). Several reports of Portlandian beds on the east coast of Kuhn Island have been based on the identification of *Aucella mosquensis* (Von Buch) (Koch, 1929a, p. 178; Frebold, 1932, p. 32; 1933, p. 62), but Maync (1940, p. 24; Vischer, 1943, p. 16, 27–29, 158) has shown that the beds in which these species of *Aucella* occur contain lower Kimmeridgian ammonites, such as *Amoeboceras* (*Amoebites*?) and *Rasenia*. The presence of *Dorsoplanites* aff. *D. panderi* (D'Orbigny) (Koch, 1935, p. 101; Frebold, 1932, p. 31) is not evidence of a Portlandian age, as Spath (1936, p. 154, 155, 170–172) records the genus in Milne Land in both upper Kimmeridgian and lower Portlandian beds.

90. The Kuhn formation occurs on the southeastern part of Kuhn Island. It consists mainly of grayish-black, friable, micaceous, pyritic sandy shale and contains the ammonites *Pavlovia*, *Dorsoplanites*, and *Pectinatites*, of upper Kimmeridgian age (Maync, 1947, p. 39, 131, 132). Similar beds of the same age probably occur on Wollaston Forland.

91. The Kløft I formation is named for a locality on the east coast of Koldewey Island, where about 71 meters of sandstone of upper Oxfordian and lower Kimmeridgian ages is exposed (Koch, 1929a, p. 167–169; 1929b, p. 254, 255; 1935, p. 99, 100; Ravn, 1911, p. 442–445; Frebold, 1932, Pls. 1–3). Equivalent beds on Kuhn Island, Wollaston Forland, and Hochstetter Forland range from a feather edge to 630 meters in thickness and consist mainly of dark-gray to black micaceous shale but contain some beds of limestone and sandstone, especially in their lower part (Vischer, 1943, p. 16; Koch, 1935, p. 100, 101; Maync, 1947, p. 128–131). Fossils from these beds have been described or determined by Ravn (1911), Rosenkrantz (Koch, 1935, p. 99, 100), and Spath (1935, p. 73, 74, 76–78). The assignment of part of the formation to a lower Kimmeridgian age is based mainly on the presence of *Rasenia*, *Aulacostephanus?*, and the subgenera *Amoebites* and *Hoplocardioceras* of the genus *Amoeboceras*. The assignment of part to an upper Oxfordian age is based on the presence of *Amoeboceras* (*Prionodoceras*) at the type locality of the Kløft I formation (Koch, 1929a, p. 168; Ravn, 1911, p. 486–488, Pl. 35, fig. 10; Pl. 36, figs. 1–3). Presumably certain forms of *Amoeboceras* (*Prionodoceras*) (Spath, 1935, p. 17, 25, 37, 74, 78, Pl. 12, fig. 4) from Koldewey Island and Wollaston Forland are also from the lower part of the Kløft I formation.

92. The Traekpas formation, named for a locality in the southern part of Koldeway Island, consists of 50 to 400 meters or more of reddish-brown to yellowish sandstone and sandy shale but includes some fine conglomerate and thin coal beds (Koch, 1929b, p. 253, 254; 1935, p. 97–99; Vischer, 1943, p. 15; Ravn, 1911, p. 441, 442). It appears to be identical with the Mauer sandstone of Frebold (1933, p. 21), named for a hill in the Wollaston Forland, and with the Yellow series of Maync (1947, p. 120–127). Its age assignment is based on the presence of *Arctocephalites*, identified by Spath (Maync, 1940, p. 22), and *Kepplerites* (Ravn, 1911, p. 489, 490, Pl. 36, figs. 5, 6; Pl. 37, fig. 1; Spath, 1932, p. 83–85). Somewhat higher in the formation occurs a pelecypod fauna that has been assigned to the Oxfordian (Maync, 1940, p. 22; Vischer, 1943, p. 15), but this age determination needs confirmation. There is as yet no positive fossil evidence of upper Callovian or Lower Oxfordian beds in Greenland. However, Frebold (1933, p. 26) suggests that the upper part of the yellow sandstone sequence is of lower Oxfordian age because it appears to grade into the shales containing *Amoeboceras*. Maync (1947, p. 127, 128) notes that the yellowish sandstones are overlain transitionally by 30 to 40 meters of soft, gray, sandy shale interbedded with shaly, gray, fine-grained sandstone, and that these gray beds grade into the overlying black shales of lower Kimmeridgian age. The gray beds contain a few cardioceratids, which show that they are not older than the Oxfordian.

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Chart 8C. - (Continued) CORRELATION OF THE

Series	European Stages (Arkell, 1946)	English Formations (After Arkell 1933)	Mexican Faunal Divisions		Cuba	Northern Central America	Southern Mexico: Tabasco, Chiapas, Eastern Veracruz	Mixteca Alta Area of western Oaxaca	Cualac Area, Northeast Guerrero	Southern Puebla	Huasteca Area of Veracruz, Puebla and Hidalgo	Tamazunchale Area of San Luis Potosi	Tampico Area of Tamaulipas and Veracruz	Northern Mexico: Sierra de Catorce, San Luis Potosi	Mountains west of Ciudad Victoria	Sierra de San Carlos, northern Tamaulipas	Mountains southeast of Saltillo	Mazapil Area of northeast Zacatecas	Sierra de Parras and Jimulco, Southern Coahuila	Villa Juarez, eastern Durango	San Pedro Del Gallo eastern Durango	Sierra San Marcos, Eastern Coahuila	San Ambrosio Well[81], Northern Nuevo Leon	Placer de Guadalupe, Eastern Chihuahua	Sonora
					65	66	67	68	69	70	71	72	73	74	75	76	77	78	79	80	81	82	83	84	85
Overlying beds					Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous		Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	
UPPER JURASSIC	Portlandian	Purbeck	Substeueroceras and Proniceras	Microcanthoceras and Hildoglochiceras	Viñales ls. 27 / Ajatychus beds / Quemado fm.			33 Limestone, dolomite and some shale			39 Pimienta formation	39 Pimienta formation				(Not identified) 48 limestone									
		Portland beds	Durangites																						
			?	Aulacosphinctoides			30 Dark limestone and shale						42 Dark limestone and shale	43 La Caja formation	46 La Casita formation		46 La Casita formation	46 La Casita formation	46 La Casita formation	46 La Casita formation	46 La Casita formation	46 La Casita formation	46 La Casita formation	46 La Casita formation	46 La Casita formation	
	Kimmeridgian	Kimmeridge clay	Mazapilites			29 Lower part of Todos Santos beds		34 Dark shale and limestone	(Not identified)	(Not identified)	40 Taman formation	40 Taman formation													
			Waagenia																						
			Idoceras duranguense and Glochiceras Tiziani																						
			Group of Idoceras balderum																						
			?																						
			Group of Sutneria platynota																						
	Oxfordian	Corallian beds	Dichotomosphinctes		Jagua 28 formation		(Not identified)	35 Limestone, marl and sandstone						44 Zuloaga limestone	Olvido fm. 47 / Zuloaga "Navilla" limestone 44		Divide fm. 47 / Zuloaga limestone 44	Zuloaga limestone 44	Zuloaga limestone 44	La Gloria 49 formation	La Gloria 49 formation	La Gloria 49 formation	La Gloria 49 formation		La Gloria 49 formation (Gypsum at top)	
			?				31 Redbeds and salt	31 Probably redbeds	31 Probably redbeds	31 Probably redbeds and salt	45 Huizachal formation (Locally absent)		45 Huizachal formation (Locally absent)	45 Huizachal formation	45 Huizachal formation					45 Huizachal formation					
	Callovian	Oxford clay	Erymnoceras mixtacorum																						
		Kellaways beds	Subgrossouvria neogaea					36 Sandstone shale and limestone	36 Dark shale and limestone	(Not identified)	36 Dark limestone					(Not exposed)								(Not identified)	
		Cornbrash beds	Eurycephalites boesai and Strenoceras paracontrarium																						
MIDDLE JURASSIC	Bathonian	Great oolite	?				32 Conglomerate and sandstone, partly red; some shale and marl with fossil plants	37 Shale, sandstone, marl, limestone, conglomerate, coal; mostly gray	38 Dark marl and limestone		32 Variegated conglomerate, shale, marl, some coal		(Not exposed)				(Not exposed)	(Not exposed)	(Not exposed)	(Not exposed)	(Not exposed)	(Not exposed)	(Not drilled)			
	Bajocian	Inferior oolite	Strenoceras aff. S. bifurcatum						38 Sandstone, coal, conglomerate																
			Teloceras																						
LOWER JURASSIC	Toarcian	Upper Lias	?					37 Black shale, some sandstone and coal			41 Huayacocotla formation														
		Middle Lias																							
	Pliensbachian	Lower Lias	Uptonia cf. U. jamesoni																						
			Microderoceras cf. M. bispinatum																						
	Sinemurian		Echioceras aff. E. raricostatum																						
			Arnioceras james-dana																						
			Pleurechioceras aff. deciduum																						
			Vermiceras aff. V. bavaricum																						
			Oxynoticeras aff. O. oxynotum																						
			Eucoassiceras aff. E. sauzeanum																						
			Arnioceras aff. A. geometricum																						
	Hettangian		Coroniceras aff. C. bisulcatum																						
Underlying beds						Permian	Permian	Schist	Schist	Schist			Intrusive igneous rocks	Partially metamorphosed sediments	Permian								Permian	Upper Triassic	

Undifferentiated conglomerate, sandstone, shale, marl, limestone, gypsum, and rock salt, probably represents most of the Jurassic

Chart 8C. - (Continued) CORRELATION OF THE JURASSIC ROCKS OF THE GULF REGION, PACIFIC COAST REGION, ALASKA, AND GREEN

By Ralph W. Imlay

Northern Mexico | Southern United States | California | Western Nevada | Oregon | Oregon | Washington

	Sierra de San Carlos, northern Tamaulipas	Mountains southeast of Saltillo	Mazapil Area of northeast Zacatecas	Sierra de Parras and Jimulco, Southern Coahuila	Villa Juarez, eastern Durango	San Pedro Del Gallo eastern Durango	Sierra San Marcos, Eastern Coahuila	San Ambrosio Well[8], Northern Nuevo Leon	Placer de Guadalupe, Eastern Chihuahua	Sonora	Malone Mountains, Western Texas	Gulf Region of United States	Coast Ranges	Sierra Nevada, west side or Gold Belt	Sierra Nevada Oroville Area, Butte County	Sierra Nevada, northern part, Downieville and Colfax Quads.	Sierra Nevada, northern part Taylorsville Area (Mt. Jura) After Diller, Hyatt and Muller	Sierra Nevada, northern part Taylorsville Area (Mt. Jura) (After Crickmay)	Klamath Mts., eastern part Redding and Lassen Peak Areas	Hawthorne and Tonopah Quadrangles	West Humboldt Range	Southwestern corner		East-central Oregon, Crook, Grant and northern Harney Counties (After Lupher)	Cascade Range northern part, north of Mt. Baker	Characteristic fossils in Oregon, northern California, and western Nevada	Characteristic fossils in Alaska
	76	77	78	79	80	81	82	83	84	85	86	87	88	89	90	91	92	93	94	95	96	97		98	99	
	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Lower Cretaceous		Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	Upper Cretaceous		Tertiary	Tertiary	Tertiary	Upper Cretaceous	Tertiary		Lower Cretaceous		Upper Cretaceous	Lower Cretaceous		
	(Not identified)										(Not identified)	Shuler formation	Knoxville formation					Combe formation				Dalhan formation	Dillard formation			Subsaueroceras	Aucellas simi
	limestone												Franciscan formation					Trail formation								Kossmatia and Durangites	Aucella piec
	La Casita formation	La Casita formation	La Casita formation	La Casita formation	La Casita formation	La Casita formation	La Casita formation	La Casita formation	La Casita formation		Malone formation with basal gypsum	Bossier fm.		(Possible intrusions)				Lucky S formation						Lonesome formation (possibly Cretaceous)		?	Aucella mosqu and Aucella rug
		Divide fm.										Cotton Valley group						Cooks Canyon formation				Galice formation	Galice group			Amoeboceras (Amoebites) and Aucella errington	Amoeboce and Aucella bro
	Zuloaga limestone	Zuloaga limestone	Zuloaga limestone	La Gloria formation	La Gloria formation	La Gloria formation	La Gloria formation		La Gloria formation (Gypsum at top)			Unnamed fm.		Slates											Shuksan formation		Cardioceras
					Huizachal formation							Smackover formation		Indian Gulch Aggls. of Taliaferro													
												Eagle Mills formation		Aguja Fria slates													
														Penyon Blanco Aggls.													
														Hunter Valley cherts													
														Pillow basalts													
														Aggls., tuffs, and flows													
	(Not exposed)																Hinchman tuff										Pseudocadoceras
																	Bicknell sandstone	North Ridge fm. Hinchman tuff Bicknell ss.									
																		Hull agglomerate									
																	Unnamed volcanics	Moonshine formation									Crenocephal and Procerites
																	Mormon sandstone	Mormon sandstone									
																	Thompson limestone	Thompson limestone Fant andesite									
																	Fant andesite	Hardgrave sandstone									
																		Lilac formation									
																	Hardgrave sandstone										
																	Trail formation										

ULF REGION, PACIFIC COAST REGION, ALASKA, AND GREENLAND

mlay

Prepared under the auspices of the
National Research Council, Carl O

	California			Western Nevada		Oregon	Oregon	Washington	
	Sierra Nevada, northern part Taylorsville Area (Mt. Jura)		Klamath Mts., eastern part Redding and Lassen Peak Areas	Hawthorne and Tonapah Quadrangles	West Humboldt Range	Southwestern corner	East-central Oregon, Crook, Grant and northern Harney Counties	Cascade Range northern part, north of Mt. Baker	Characteristic fossils in Oregon, northern California, and western Nevada
	After Diller, Hyatt and Muller	(After Crickmay)				Diller / Tallia-ferro	(After Lupher)		
91	92	93	94	95	96	97	98	99	
Tertiary	Tertiary	Tertiary	Upper Cretaceous	Tertiary		Lower Cretaceous	Upper Cretaceous	Lower Cretaceous	
		Combe formation 60				Dothan formation 63 / Dillard formation 63			Substeueroceras
		Trail formation							Kossmatia and Durangites
		Lucky S formation							
		Cooks Canyon formation							
	60 Foreman formation	Foreman formation				Galice formation 63 / Galice group 63	64 Lonesome formation (possibly Cretaceous)	65 Shuksan formation	Amoeboceras (Amoebites) and Aucella erringtoni
57									
	Hinchman tuff								?
	Bicknell sandstone	North Ridge fm. / Nicolaus fm. / Bicknell ss.			(Not identified)	Dothan group 63	Trowbridge shale		Lillooettia and Cosmoceratidae
		Hull agglomerate							
	Unnamed volcanics	Moonshine formation							Staphanoceratidae and Sonninites
59	Mormon sandstone	Mormon sandstone				58 Plant beds of Douglas County	Snowshoe fm. / Hyde fm.		
	Thompson limestone	Thompson limestone / Fant andesite					Colpite group: Warm Springs formation / Weberg formation		Sonninitidae and Sphaeroceratidae
	Fant andesite	Hardgrave sandstone	61 Potem formation	62 Dunlap formation					Timetoceras and Sphaeroceratidae
		Lilac formation					Mowich group: Nicely shale / Suplee formation / Robertson formation		?
	Hardgrave sandstone		Begley andesite						Amaltheus, Harpoceras, Amauroceras and Hildoceratidae
							Donovan formation		Plicatostylus
	Trail formation		Modin formation	Sunrise formation	62 Limestone of Muttlaberry Canyon				Eoderoceras
									?
									Coroniceras and Arniceras
									?
									Euphyllites and Pelloceras
Jassic	Upper Triassic	Upper Triassic	Upper Triassic	Upper Triassic	Upper Triassic		Triassic	Triassic	

	Alaska														East Gre
Characteristic fossils in Alaska	Alaska Peninsula		Cook Inlet			Alaska Range (Yenta and Skwenta Valleys)	Matanuska Valley and Talkeetna Mtns.	Copper River Region (Chitina Valley, and Nutzotin Mtns.)	Southeastern Alaska (Juneau District)	Lower Kuskokwim Region	Northern Alaska north of Brooks Range	Characteristic fossils in East Greenland Modified after Spath (1932, 1935, 1936)	Milne Land	Jameson Land	Tr / Ge / Soc
	Wide Bay	Puale Bay	Seldovia–Port Graham	Tuxedni Bay	Iniskin Peninsula										
100	100	101	102	103	104	105	106	107	108	109	110		111	112	
				Tertiary	Tertiary		Lower Cretaceous	Lower Cretaceous		Lower Cretaceous	Upper Cretaceous			Lower Cretaceous	Cr
Aucellas similar to Aucella piochii	(Not identified)	(Not identified)		(Not identified)	(Not identified)		(Not identified)	76	79 Dark shale			Titanites? and Craspedites Lougaites	? 82 Hartzfjaeld sandstone	83	
												?			
												Crendonites beds	Sandy sh.		
												Epipallasiceras beds	? Glauconitic series		
												Pallasiceras beds	Sandy clay		
Aucella mosquensis and Aucella rugosa	66	66		66	66		66			80	81	Pectinatites beds	Reddish sandy marl		
										Present					
	Naknek formation	Naknek formation		Naknek formation	Naknek formation		Naknek formation	Slate, graywacke, and conglomerate with Aucella	? 79 Douglas Island volcanic group			?	Dark shales		
Amoeboceras and Aucella bronni						(Not identified)						Hoplocardioceras ilasipinense / Euprionoceras kochi / Rasenia borealis / Rasenia orbignyi	Amoebites shales	84 Koch Fjaeld formation (Klaft I fm.)	
									Treadwell slate			Amoeboceras (Prionodoceras) and Ringsteadia	Pecten sandstone		
Cardioceras												Cardioceras aff. C. tenuidos	Cardioceras shales		
	(Not identified)	(Not identified)		Chitk congl. (Not identified)	Chitk congl. (Not identified)							Cardioceras beds	Charcot Bay ss.		
	67	67		70	70		70					?			
Pseudocadoceras	Shelikof formation	Shelikof formation		Chinitna formation	Chinitna formation		Chinitna formation	?							
												Kepplerites tychonis beds		85	
				(Not identified faunally)	(Not identified faunally)							Arcticoceras beds		Vardekloft formation (Fossil Mountain formation)	
Cranocephalites and Procerites (?)				71 Bowser member	71 Bowser member	73 Tordrillo formation	71 Tuxedni formation	77 Tuffaceous slate near mouth of Chitina River				Arctocephalites beds			
Defonticeras				Cynthia Falls ss. member / Siltstone mem.	Cynthia Falls ss. member / Siltstone mem.							Cranocephalites beds		(Not identified)	
Emilaia				Sandstone and Siltstone	(Present in subsurface)				79 Thane volcanic group			?			
Timetoceras	68 Kialagvik formation														
		69		72	72	(Not identified)	74 Talkeetna formation	78 Tuffaceous conglomerate near Taral				Pseudolioceras / Dactylioceras		86 Nails Cliff formation	
Pseudogrammoceras Dactylioceras and Peronoceras												?			
Amaltheus Deroceras and Xipheroceras															
	(Not exposed)	(Present)		(Probably represented by tuffs, lavas and agglomerates)	(Probably present in subsurface and in Marsh Creek Area)	Skwenta group	Not identified					Beaniceras / Uptonia jamesoni			
Coroniceras and Arniceras												?			
Wahneroceras (?)												Plant beds			
	Upper Triassic	Upper Triassic		Upper Triassic	Upper Triassic	Devonian	Carboniferous		Upper Triassic	Upper Triassic	Upper Triassic		Granite	Cape Stewart fm. (upper part) Upper Triassic (?)	

FIC COAST REGION, ALASKA, AND GREENLAND

Prepared under the auspices of the Committee on Stratigraphy of the National Research Council, Carl O. Dunbar, *Chairman*

Western Nevada — Oregon — Oregon — Washington

Alaska — East Greenland

Western Nevada: Hawthorne and Tonopah Quadrangles	Western Nevada: West Humboldt Range	Oregon: Southwestern corner, Dillar	Oregon: Southwestern corner, Tallaferro	Oregon: East-central Oregon, Crook, Grant and northern Harney Counties (After Lupher)	Washington: Cascade Range northern part, north of Mt. Baker	Characteristic fossils in Oregon, northern California, and western Nevada	Characteristic fossils in Alaska	Alaska Peninsula: Wide Bay	Alaska Peninsula: Puale Bay	Cook Inlet: Seldovia–Port Graham	Cook Inlet: Tuxedni Bay	Cook Inlet: Iniskin Peninsula	Alaska Range (Yenta and Skwenta Valleys)	Matanuska Valley and Talkeetna Mtns.	Copper River Region (Chitina Valley, and Nutzotin Mtns.)	Southeastern Alaska (Juneau District)	Lower Kuskokwim Region	Northern Alaska north of Brooks Range	Characteristic fossils in East Greenland Modified after Spath (1932, 1935, 1936)	East Greenland: Milne Land	East Greenland: Jameson Land	East Greenland: Traill Island and Geographical Society Island	Northern Area: Kuhn Island, Kaldevejy Island, Wollaston Foreland, Clavering Island, Hochstetter Foreland	Northwest Europe Standard Zones (Arkell, 1946)
95	96	97		98	99			100	101	102	103	104	105	106	107	108	109	110		111	112	113	114	
Tertiary		Lower Cretaceous		Upper Cretaceous	Lower Cretaceous						Tertiary	Tertiary		Lower Cretaceous	Lower Cretaceous		Lower Cretaceous	Upper Cretaceous			Lower Cretaceous	Lower Cretaceous	Lower Cretaceous	
		63 Dothan formation	63 Dillard formation	64 Lonesome formation (possibly Cretaceous)		Substeueroceras; Kossmatia and Durangites	Aucellas similar to Aucella piochii	(Not identified)	(Not identified)		(Not identified)	(Not identified)		(Not identified)	76	79 Dark shale	80 Present	81	Titanites? and Crespedites Lougeites	? 82 Hartz fjeld sandstone	83	88	89 (Present on Kühn Island) (Rigi series)	Titanites gigenteus; Korberites okusensis
							?												?					Glaucolithites goni
							Aucella mosquensis and Aucella rugosa	66	66		66	66		66					Crendonites beds	Sandy sh. ? Glauconitic series ? Sandy clay	(Possibly present)		Kühn formation 90	Zaraiskites alboni; Paviavia pallasioides
																			Epipallasiceras beds	Reddish sandy marl				Paviavia rotunda
							?												Pallasiceras beds	Dark shales				Pectinatites pectinatus
																			Pectinatites beds					Subplanites wheatleyensis; Subplanites sp.
																			?				(Not identified)	Gravesia gigas; Gravesia gravesiana
							Amoeboceras and Aucella rugosa												Hoplocardioceras decipiens; Euprionoceras kochi; Rasenia borealis; Rasenia orbignyi	Amoebites shales	? 84 Koch Fjeald formation (Klafft I fm.)		? 91 Klafft I formation (black series) (dark shale containing beds of limestone and sandstone)	Autocosphinctes; Rasenia mutabilis; Rasenia cymodoce; Pictonia baylei; Ringsteadia pseudocordata; Decipia decipiens
							?												Amoeboceras (Prionodoceras) and Ringsteadia	Pecten sandstone				Periphinctes antiflexiplex; Periphinctes plicatilis
																			Cardioceras aff. C. tenuidos	Cardioceras shales				Cardioceras cordatum; Quenstedtoceras mariae; Quenstedtoceras lamberti; Peltoceras athleta
							Cardioceras												Cardioceras beds	Charcot Bay ss.			Possibly represented by gray shale and sandstone	Erymnoceras coronatum; Kosmoceras jason; Sigaloceras calloviense; Proplanulites koenigi; Macrocephalites macrocephalus; Clydoniceras discus
																			?					(not yet determined)
							Pseudocadoceras; Parasidoceras, Pleurites, Litetia, Thamnophyllites, Sutneria, Cadoceras, Gowericeras	(Not identified) 67 Shelikof formation	(Not identified) 67 Shelikof formation		(Not identified) 70 Chinitna formation	(Not identified) 70 Chinitna formation		70 Chinitna formation	?			Hoplites tychonis beds		85 Vardekløft formation (Fossil Mountain formation)	88 (Present)	92 Traekpas formation (Yellow sandstone and conglomerate) (yellow series)	Parkinsonia parkinsoni; Stephanoceras humphriesianum
																			Arcticoceras beds					Otoites sauzei; Sonninia sowerbyi; Ludwigia murchisonae; Lioceras opalinum
							Cranocephalites and Procerites (?)				71 Tuxedni formation: Bowser member (Cynthia Falls ss. member; Siltstone mem. ss., siltstone, gray siltstone, reddish brown weathering; Sandstone and Siltstone) 72 (Probably represented by tuffs, lavas and agglomerates)	71 Tuxedni fm.: Bowser member (Cynthia Falls ss. member; Siltstone mem.; Siltstone mem.) (Present in subsurface) 72 (Probably present in subsurface and in Marsh Creek Area)		71 Tuxedni formation	73 Tardrilla formation	77 Tuffaceous slate near mouth of Chitina River		Arctocephalites beds					Lytoceras jurense; Hildoceras bifrons; Harpoceras serpentinum; Dactylioceras tenuicostatum; Polyplectus spinatum; Amaltheus margaritatus; Prodactylioceras davoei; Tragophylloceras ibex; Uptonia jamesoni; Echioceras raricostatum; Oxynoticeras oxynotum; Asteroceras obtusum; Arietites furneri; Arnioceras semicostatum; Coroniceras buckiandi; Scamnoceras angulatum; Psiloceras planorbis
							Eoderoceras												Cranocephalites beds					
							?												?					
							Stephanoceratidae and Sonninites; Sonniniidae and Sphaeroceratidae; Tmetoceras and Sphaeroceratidae	Emileia; Tmetoceras													Pseudolioceras; Dactylioceras					
							?												?					
							Amaltheus, Harpoceras, Amauroceras and Hildoceratidae	Pseudogrammoceras Dactylioceras and Peronoceras											Beaniceras; Uptonia jamesoni					
							Plicatostylus	Amaltheus Deroceras and Xipheroceras											?					
							Eoderoceras												Plant beds					
							?	Coroniceras and Arnioceras																
							Coroniceras and Arnioceras																	
							?	Wahneroceras (?)																
							Euphylites and Psiloceras																	
Upper Triassic	Upper Triassic	Triassic		Triassic	Triassic				(Not exposed)	Upper Triassic	Upper Triassic	Upper Triassic	Devonian	Carboniferous		Upper Triassic	Upper Triassic	Upper Triassic		Granite	Upper Triassic (?)	Upper Triassic	Paleozoic	

Western Nevada: 61 ? ion; 62 Dunlap formation; ? Sunrise formation; Upper Triassic
Western Nevada: 62 Limestone of Muttleberry Canyon; Upper Triassic
Oregon: 58 Plant beds of Douglas County; Upper Triassic
Oregon: Coburg limestone member; Snowshoe fm.; Hyde fm.; Worm Springs formation; Weberg formation; Nicely shale; Suplee formation; Robertson formation; Mowich group; ? Donovan formation; ? Triassic
Oregon: Galice formation; Galice group; 63 Dothan group; ? Trowbridge shale; ? Triassic
Oregon: 64 Lonesome formation (possibly Cretaceous); ? Triassic
Washington: ? Triassic
Washington: 65 Shuksan formation; ? Triassic
Alaska Peninsula: (Not exposed); Upper Triassic
Cook Inlet: ? 68 Kialagvik formation; ? 69 (Present); Upper Triassic
Cook Inlet: ? 72 (Present); ? Upper Triassic
Cook Inlet: Upper Triassic
Alaska Range: (Not identified); Devonian
Matanuska Valley: ? 74 Skwenta group; Not identified; Carboniferous
Copper River Region: ? 75 Talkeetna formation; ? 78 Tuffaceous conglomerate near Taral; Carboniferous
Southeastern Alaska: ? 79 Douglas Island volcanic group; Treadwell slate; ? Thane volcanic group; Upper Triassic
Lower Kuskokwim Region: Kingak shale; Upper Triassic
Northern Alaska north of Brooks Range: Upper Triassic
East Greenland: Granite
East Greenland: Cape Stewart fm. (upper part) 87; Upper Triassic (?); 86 Neills Cliff formation; ? Cape Stewart fm. (lower part); Upper Triassic
East Greenland: (Not identified); Upper Triassic